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CORRECTIONS AND EMENDATIONS

Contributors have been invited to send corrections and emendations to be made in their papers, and the volume has been scanned with some care. Corrections and insertions are as follows:

Page 1148, line 24, *for* time, *read* distance.
" 1252, line 16, *for* underload, *read* under load.
" 1253, line 20, *for* comprehensive, *read* compressive.
" 1352, line 22, *for* Micronichnus, *read* Micrichnus.
" 1409, line 12 from bottom, *insert* (4) *before* Weathering
" 1438, column 5, "Buckner red beds and anhydrite" belongs in box directly below
" 1813, line 26, *delete* Geological Society project.

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ENDOMORPHISM OF THE IDAHO BATHOLITH

BY ALFRED L. ANDERSON

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ABSTRACT

Study of many parts of the Idaho batholith shows considerable endomorphism by post-consolidation solutions. The solutions, enriched particularly in silica but also in variable but generally appreciable quantities of potash and minor amounts of other elements, produced many changes in the solid rock. Biotite, orthoclase, microcline, muscovite, quartz, sphene, apatite, zircon, magnetite, garnet, allanite, epidote, zoisite, chlorite, and sericite replaced earlier minerals. As a result, the final rock is considerably more silicic and generally more alkalic

Much of the dioritic marginal rock of the batholith has been changed to a quartz-rich diorite (tonalite) and granodiorite, and the less calcic rock of the interior from a quartz-bearing diorite and granodiorite to a somewhat calcic quartz monzonite with variations from granodiorite to granite. Dikes of associated aplite show similar end-stage changes, but most of the pegmatites have formed from somewhat later hydrothermal solutions by direct replacement of the aplitic and batholithic rock along fractures.

INTRODUCTION

For some time evidence has been accumulating that the rock of the Idaho batholith and adjacent similar masses (Fig. 1) has been considerably modified by post-consolidation emanations and that much of the intrusive masses is more silicic and alkalic than the original magma. Such modifications of the already consolidated rock were first noticed in granodiorite in the vicinity of Pend Oreille Lake in northern Idaho (Gillson, 1927, p. 1-20) and later over a much wider area in that region (Anderson, 1930a, p. 29-30; 1940, p. 17-21) and in the rock of the Cassia batholith in the extreme south of the State (Anderson, 1934a, p. 376-392). Mention of any such changes in the rock of the Idaho batholith, however, did not appear until 1934 (Shenon and Reed, p. 16), and no detailed descriptions until 1935 (Currier, p. 7-11) and 1940 (Anderson and Hammerand, 1940, p. 561-589), each report covering a small and widely separated part of the batholith. Anderson (1934b; 1939) also reported on other parts of the batholith but merely summarized the effects of the post-consolidation emanations.

The writer's interest in the consolidation of granitic rocks in Idaho began in the Clark Fork district near Pend Oreille Lake in 1927. The rocks had been considerably modified by post-consolidation emanations (Anderson, 1930a, p. 29-30), as in the near-by Pend Oreille district (Gillison, 1927, p. 1-20). Question arose as to whether other granitic masses, including the Idaho batholith, may not have been similarly affected. As field work in the State continued, the study was extended to the Idaho batholith and other granitic masses. Work on the batholith began

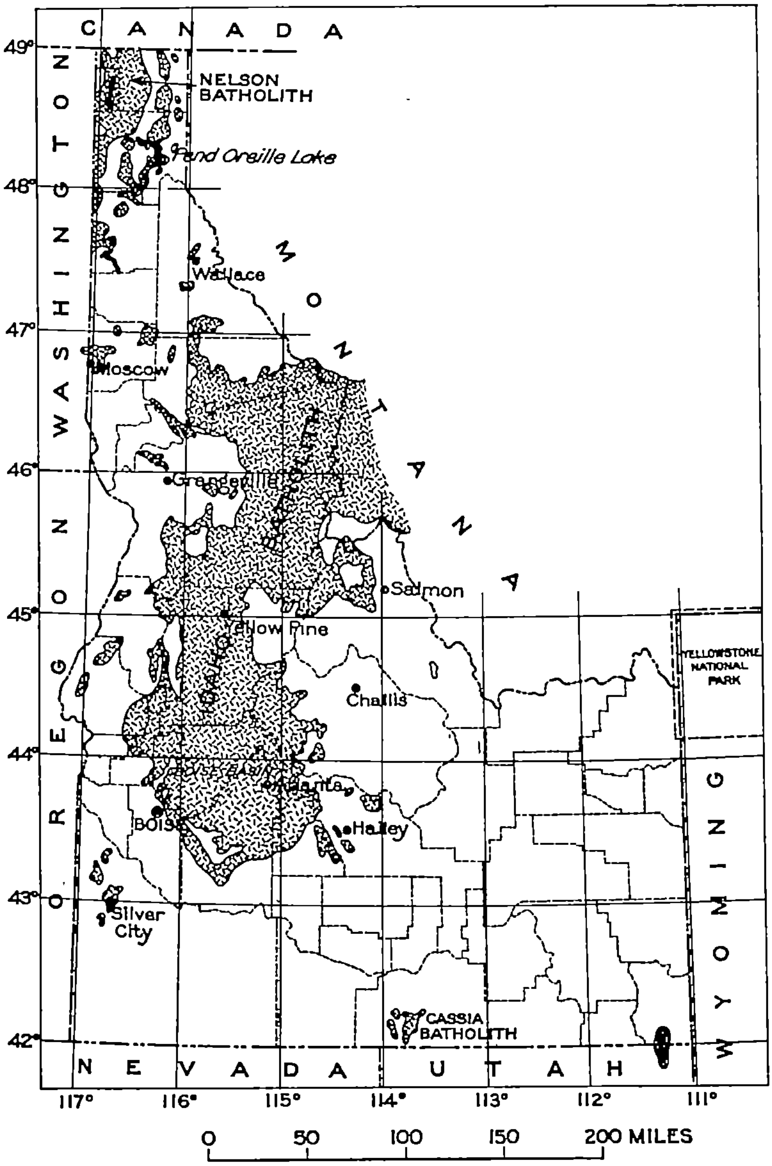

FIGURE 1.—*Sketch map showing distribution of Idaho batholith and other granitic rocks in the state*

in 1929 with examination of several thousand square miles of its northern part (Anderson, 1930b, p. 17-29). In 1930 the writer studied in detail the comparatively small Cassia batholith in southern Idaho, giving especial attention to effects of the igneous emanations on the invaded rocks (Anderson, 1934a, p. 376-392). Since 1931 much of the work on the batholith has been carried on in the south and southwest parts of the State, particularly in the Pearl-Horseshoe Bend (Anderson, 1934a, p. 1-36), Boise Basin (Anderson, 1942a), Atlanta (Anderson, 1939), and Rocky Bar (Anderson, 1942b) mining districts, but a small part of the batholith in the north part of the State (Anderson and Hammerand, 1940, p. 561-589) was also studied in detail.

Critical examination of many thin sections from all these districts as well as from other widely scattered localities in the Idaho batholith has made possible a fairly comprehensive analysis of the consolidation history of the entire batholith. Evidence from other parts indicates that the findings are applicable to the whole batholith.

In this paper only the endomorphism of the main granitic facies of the batholith and of associated aplite and pegmatite dikes is considered. The addition of igneous material to the invaded sedimentary rocks is not discussed as was done with the Cassia batholith (Anderson, 1934a, p. 376-392) and on a small part of the Idaho batholith in the north part of the State (Anderson and Hammerand, 1940, p. 561-589). Such studies in northern Idaho reveal that there is little material difference in the changes involved in progressive "granitization" of the invaded sedimentary rocks and in some of the changes induced in the granitic rock by endormorphic processes (Anderson, 1940, p. 17-20; Anderson and Hammerand, 1940, p. 561-589). However, the possible influence of incorporation of sedimentary material into at least the outer zones of the batholith cannot be wholly ignored in such speculations on the composition of the rock as are made herein.

The writer wishes to acknowledge his appreciation to G. F. Loughlin, C. S. Ross, and C. P. Ross for their assistance in the preparation of the manuscript and also the helpful suggestions made by J. C. Reed, P. J. Shenon, A H. Koschmann, and others.

DIVISIONS OF THE BATHOLITH

Although the Idaho batholith, recognized as one of the major batholiths in North America, presents many interesting relations, this paper attempts to point out merely the broader structural and compositional features and to emphasize its natural division into two main facies differing from one another in age, structure, and composition. One facies is represented in a broad marginal and roof zone of more calcic rock consolidated under considerable stress; the other is a less calcic inner zone which consolidated when stresses were less intense. In some localities more calcic rocks fringe the batholith, apparently intruded ahead of the main marginal mass, perhaps as injections of early little differentiated magma, but although the relations locally are complex, the presence of these subordinate masses does not rule out the picture of the batholith as consisting dominantly of two distinctive facies. This natural division into two main facies was first recognized as a major structural feature in the north part of the State (Anderson, 1930a, p. 17-29) and later in the southwest part where the distinction was based more particularly on differences in color and composition (Anderson and Rasor, 1933, p. 287-289). Since then others have noted the division; Ross, in the Casto quadrangle along a part of the east margin of the batholith (1934, p. 35-45), also later as a general feature (1936, p. 375-376); Currier, in the Yellow Pine district in the central part of the State (1935, p. 7-10); Shenon and Ross, in the Edwardsburg district also near the central part of the State (1936, p. 8-10); and Reed, in the Florence mining district near the west-central part of the batholith (1939, p. 7-8). In most places the two facies are transitional, but in some places the marginal rock is cut by the inner facies.

The marginal and inner facies differ appreciably in appearance and composition, and, in general, each may be readily identified. Locally the marginal facies is only a few hundred feet wide, but in most places it is thousands of feet or even miles across. The marginal facies makes up thousands of square miles of the exposed part of the batholith and equals roughly half the mass of the batholith. Because of the absence of systematic mapping, it is not feasible to show its distribution even approximately.

In places the marginal facies merges with sedimentary rocks that have undergone contact metamorphism, and it is difficult to tell how much of the marginal facies has a hybrid origin. Much of the rock discussed in this paper, however, appears to represent consolidated magma, and discussion of "granitized" borders is withheld.

MARGINAL FACIES

COMPOSITION

Much of the marginal facies is quartz diorite, but in broad areas it is granodiorite and locally quartz monzonite and granite. More calcic border rocks include peridotites, pyroxenites, gabbros, norites, and diorites, but since these escaped the general endomorphism of the main marginal facies, they are not considered further. The marginal facies is more calcic in the northwest part of the batholith, where over several thousand square miles it is quartz diorite (Anderson, 1930b, p. 18). In the south and southwest part much of it is granodiorite (Anderson and Rasor, 1933, p. 287-294). In other parts of the batholith either or both types may prevail. Both quartz diorite and granodiorite are present in the Pearl-Horseshoe Bend region (Anderson, 1934b, p. 5-6). Granodiorite largely predominates in the Atlanta district (Anderson, 1939, p. 6-7), but quartz diorite also appears at Yellow Pine (Currier, 1935, p. 7-11), in the Casto quadrangle (Ross, 1934, p. 35-38), and in the Edwardsburg district (Shenon and Ross, 1936, p. 8-9). Quartz diorite predominates in the Elk City (Shenon and Reed, 1934, p. 12-13) and Florence (Reed, 1939, p. 7-8) areas. The variations from quartz diorite to granodiorite can, in general, be detected only microscopically and consequently are not mappable.

ROCK DESCRIPTION

Most of the marginal facies is moderately dark gray and somewhat gneissic. The light-colored grains commonly range from 4 to 7 millimeters in diameter, and the dark ones from 2 to 3 millimeters. Commonly 45 to 60 per cent consists of sodic andesine, averaging about $Ab_{65}An_{35}$. (In the inner facies the plagioclase is oligoclase.) The more calcic plagioclase in the marginal facies is generally distinctive. Quartz is everywhere conspicuous. In the northwest part of the batholith, it makes up 15 to 20 per cent, but in the southern and southwestern parts it is 25 to 30 per cent. Biotite and hornblende compose 10 to 15 per cent, rarely more. Biotite is commonly the more abundant, but hornblende is the principal dark constituent locally along the northwest margin. Small amounts of microcline are usually present, excepting in the northwest margin. Locally in the south the microcline increases up to 15 per cent, and even 30 per cent. Liberal sprinkling of grains of sphene and epidote is one of the distinctive features of the facies. Minor amounts of apatite, magnetite, zircon, allanite, zoisite, chlorite, muscovite, and sericite also are present.

MINERAL RELATIONS

Instead of the usual admixture of euhedral, subhedral, and anhedral forms, resulting from magmatic crystallization, many grains irregularly penetrate or impress one another in such a way that crystal faces are the exception but recognizable if the complete grain be restored by masking out the penetrating mineral.

Except for uneven penetration by biotite and quartz, hornblende tends to show crystal outline or outline modified by mutual interference during growth. Its outline is especially uneven along quartz contacts and appears as though indented by the quartz. If the quartz indentations were filled with hornblende, however, the hornblende grain would appear in crystal outline.

The andesine also tends to have crystal outline, but because of its dominance over hornblende it has been much more affected by mutual crowding and is subhedral. However, contacts with microcline and quartz are extremely irregular; the microcline and quartz grains extend into and in places completely penetrate the andesine,—as they do the oligoclase in the inner facies (Pl. 2, fig. 3, 6; Pl. 3, fig. 1, 2, 3). Larger microcline grains enclose irregular remnants of andesine (Pl. 1, fig. 5). Although most andesine grains are clear and unzoned, some hold irregular ghostlike corroded inclusions of a more calcic, zoned plagioclase (Pl. 1, fig. 1). The calcic inclusions are rather widely scattered and appear to have been inherited, but from sources not determined. These zoned inclusions are ordinarily more highly sericitized than the host.

In general the biotite molds upon or lies between andesine and hornblende, but its shape does not conform altogether with the intergranular space. Some grains are unevenly penetrated by andesine and parts are retained in the andesine as small remnant oriented inclusions (Pl. 1, fig. 2). In some cases the penetrating andesine has fringelike borders as in Figure 2 of Plate 1, but more generally it impresses its own crystal outline against the biotite. On the other hand, some biotite grains penetrate the andesine, either maintaining essentially euhedral outlines (Pl. 1, fig. 3) or sending narrow protuberances into the andesine. Some grains penetrate one or more andesine crystals and some retain small oriented inclusions of the penetrated andesine. Thus the biotite has two habits, dependent on age, and is partly older and partly younger than andesine. Commonly the younger forms an added growth on the older (Anderson and Hammerand, 1940, p. 570). Borders against microcline and quartz are more regular; even crystal faces may show against the microcline. Against quartz, however, its borders, though smooth, are unevenly indented and grains may be cut entirely through (Pl. 1, fig. 4).

Microcline, if present in very scant amounts, generally forms roughly wedge-shaped grains among the andesine crystals; if more abundant, its grains become large and irregular. Deep embayments and irregular veinlets may then extend partly or entirely through the andesine crystals,—as they do in the oligoclase of the inner facies (Pl. 2, fig. 3). Reaction between microcline and andesine is suggested in many places by intervening zones in the andesine of differing extinction angles and composition. Wherever the microcline is more abundant and in still larger grains, it engulfs one or more andesine crystals as residual, corroded grains (Pl. 1, fig. 5), often as ghostlike shadows (Pl. 2, fig. 5). Crystal faces of hornblende and biotite determine the borders of the microcline. Euhedral biotite or hornblende is not uncommon as inclusions in microcline. Its relations to the quartz are varied. In some specimens the microcline holds widely scattered, rounded, and irregularly corroded quartz grains,—similar to those appearing in the orthoclase of the inner facies—(Pl. 2, fig. 4) which appear to be remnants of anhedra originally wedged between plagioclase laths. Such quartz inclusions are exceptional, as most of the quartz is more or less lobate in and against the microcline (Pl. 3, fig. 2).

Quartz occurs also as small globular grains in andesine that has been corroded by microcline thus simulating myrmekitic intergrowths. Such intergrowths are quite prevalent throughout the facies. The largest part of the quartz occurs as irregular seams in feldspars and dark minerals or appears as large lobate grains which indent them. Except for the embayments into the plagioclase, hornblende and biotite grains, the quartz otherwise is largely interstitial. These relations are expressed in Figure 4 of Plate 1, which shows a seam of quartz cutting through a grain of biotite, also in Figure 6 of Plate 2 and in Figures 1, 2, and 3 of Plate 3. Except for Figure 4 of Plate 1, the photographs represent the inner facies, but the relations in the marginal rock are the same. Quartz seams of two ages are shown in Figure 3 of Plate 3, one cutting across and also embaying the other. Irregular remnants and lobes of one in the other are common. Thus two late stages of quartz deposition are indicated, which, added to the earlier two, make four in all, each with its own particular habit. The last two have like characteristics, except for age difference.

Much of the sphene occurs as large rhombic crystals cutting across or protruding into and through the hornblende, biotite, and andesine in a manner that does not accord with minerals crystallized in a pyrogenic sequence. The haphazard distribution of the sphene is partly controlled by grain contacts of other minerals (Pl. 1, fig. 6) and partly by mineral cleavages. Some of the sphene also occurs as large irregular grains, some of which have tongues projecting into adjacent grains, particularly into andesine (Pl. 2, fig. 1). Its habit therefore is largely penetrative, like much of the quartz. However, its relations to the quartz are indeterminate. It neither penetrates nor is penetrated by the quartz.

Grains and crystals of allanite, like sphene, penetrate the andesine and the biotite in a haphazard manner along grain contacts and mineral cleavages. Because it is less abundant than either the sphene or the epidote, its relations are not so clearly defined. Although usually occurring as independent crystals and grains, some is mantled or enclosed by epidote.

Scattered grains of epidote as large as the leading minerals are usually extremely irregular and thread through the biotite and hornblende crystals and into the andesine (Pl. 2, fig. 2). In some places the epidote mantles grains of allanite, but most of it occurs as independent grains. Except for its irregular outlines, its relations resemble those of the sphene. Grains of zoisite appear more rarely scattered through the biotite and andesine.

Except for differences in size and amount, the grain relations of the scant accessories,—magnetite, apatite, and zircon—are like those of the sphene and allanite. The three minerals ordinarily occur in well-shaped, though small, crystals, aligned for the most part along borders of quartz, feldspar, and biotite grains, or scattered along fractures and cleavages in the feldspar or biotite.

Small grains of muscovite are widely scattered in the rock that contains large amounts of microcline. The grains ordinarily project into the andesine, less commonly into the microcline, and are closely associated with seams of quartz though contained in the bordering feldspar. Although the muscovite grains are relatively small, they are too large to be sericite and are easily distinguished from the minute grains of sericitic mica which are also common though not abundant. Much of the sericite occurs as small grains or aggregates of small grains in scattered patches in andesine. A little is found in thin seams which extend into and through the quartz grains.

Chlorite is not characteristic of the facies and is nowhere conspicuous in the fresh rock. Much of the chlorite represents partly altered biotite. It is subordinate and scarce as compared to epidote.

INTERPRETATION OF MINERAL RELATIONS

The mineral grains are intricately penetrated or irregularly penetrate neighboring grains so that their relations are not typical of orderly crystallization but resemble those of minerals formed by replacement found in alteration zones in wall rock bordering high-temperature ore deposits. Were most of the irregular penetrating grains masked by filling the embayments with the substance of the penetrated grains, euhedral, subhedral and interstitial auhedral grains would appear.

The hornblende does not have good crystal outline because it is penetrated by some of the grains of biotite, sphene, epidote, and quartz; and the andesine would be subhedral were its borders not so irregularly indented by biotite, microcline, and quartz. Some of the hornblende shows crystal faces against the andesine, and some andesine has crystal faces next to scattered wedge-shaped interstitial microcline and quartz grains. These interstitial grains are those that appear as inclusions in the larger grains of microcline. In general the biotite lies along and between the andesine, some of it penetrated by andesine and some of it penetrating andesine or sending narrow protuberances into the andesine. Even though the sphene, allanite, magnetite, apatite, and zircon are euhedral in general they seem to be penetrating minerals.

On restoring the original crystal outlines the relations of the hornblende, andesine, and some of the biotite, potash feldspar, and quartz are those in ordinary consolidated magma.

Modification was probably orderly; minerals were added in rather definite sequence. The irregular protuberances from the biotite into and through the hornblende and andesine crystals suggest an added growth to grains originally between the hornblende and andesine crystals by replacement of the adjacent minerals. Late formation of biotite by replacement has been described by Gillson (1929, p. 77-86) in endomorphism in the Pioche district and by Anderson and Hammerand (1940) in a part of the Idaho batholith. Biotite is known to form in contact metamorphic zones in hydrothermal assemblages.

The microcline also indicates much replacement. Its extensive penetration of the andesine, accompanied in places by reaction rims of more silicic zones, and the presence in the microcline of numerous oriented remnants and shadowlike traces of andesine cannot be interpreted otherwise. On similar evidence Gillson (1927, p. 8) and the writer (1930a, p. 29-32) ascribed a replacement origin for the microcline in the granitoid rocks near Pend Oreille Lake; Currier (1935, p. 8-10) regarded the microcline in the Yellow Pine district as being added during the late stages of consolidation. In the Elk City region Shenon and Reed (1934, p. 16) pointed out such relations evidenced by reaction between microcline and plagioclase. Replacement of quartzite by microcline occurs in the contact zone of the Cassia batholith (1934a, p. 376-392) and the Idaho batholith in northern Idaho (Anderson and Hammerand, 1940, p. 575-579). The microcline shows no penetration nor corrosion of biotite but locally contains perfect crystals as inclusions, so it probably formed shortly after the late biotite and some of the pyrogenic quartz, retained as rounded and embayed inclusions.

The relations of much of the quartz also suggest its introduction by replacement. A minor amount has penetrated and replaced the andesine in myrmekiticlike blebs, but most of it occurs in large lobate grains, in seams, and in veinlets which penetrate the andesine, hornblende, and biotite, and in places the microcline. Its replacement relations seem to be identical to those of the quartz in the Sparta granite. Gilluly (1933, p. 65-81) emphasized the lobate habit of the quartz as one of the best criteria of replacement in the Sparta granite. Currier (1935, p. 8) regards the late microcline and quartz in the Yellow Pine district as contemporaneous, but the writer has observed that in some sections the lobate quartz partly penetrates the margins of the microcline grains (Pl. 3, fig. 2). Seams and lobes of quartz are invaded by other seams and lobes of quartz, so that the late quartz is not all the same age.

The manner in which the sphene protrudes into and extends across grains of hornblende, biotite, and andesine, and its position in part controlled by fracturing and by cleavages, suggests replacement. Although irregular tongues may extend into and replace adjoining grains of andesine, the sphene commonly assumes its own crystal form during the replacement. Thus in its haphazard arrangement and other relations it is identical to the sphene in the granodiorite in the Pend Oreille region, which Gillson (1927, p. 5-8, 27-29) describes as a late introduction and replacement. The sphene in the granites in the Yellow Pine district is similarly regarded by Currier (1935, p. 8-11).

Much of the magnetite, apatite, and zircon is in small crystals concentrated along grain contacts and in fractures and cleavages of the earlier minerals, so that in occurrence they resembled the sphene. Gillson (1927, p. 16) pointed out these relations for the Pend Oreille district and Currier (1935, p. 8, 10) for the Yellow Pine district. They concluded that the minerals were added during a late replacement sequence, and the writer concurs.

Other minerals not already discussed likewise appear to be late replacements. The allanite has a distribution and habit not unlike that of the sphene and, as interpretated by Gillson (1927, p. 8, 27-29), has formed at the expense of the earlier minerals. The epidote occurs generally as large irregular grains with penetrating apophyses in and across cleavages of invaded minerals, and it, too, apparently is a late replacement similar to the late epidote described by Gillson (1927, p. 16-20). Its mantling of some of the allanite proves that the epidote is somewhat later. Chlorite occurs only as a minor alteration product of biotite in fresh rock, suggesting a late place in the sequence, as in the Pend Oreille district (Gillson, 1927, p. 16-20). The relations of the muscovite suggest replacement of the feldspars, possibly closely succeeding the formation of the microcline, but, as pointed out by Currier (1935, p. 8-10), as a phase of a still later stage in which sericite was formed. Narrow seams of sericite cutting the late quartz indicate its very late origin.

In addition to the sequence of minerals formed by crystallization another sequence is apparently superposed on the first by replacement. Thus to the succession of minerals of the early consolidated rock—the hornblende, biotite, andesine, in places a little potash feldspar, and generally scant amounts of quartz—are added the younger replacement sequence: (1) biotite, microcline, and quartz; (2) sphene, zircon, apatite, and magnetite; (3) quartz, allanite, epidote, chlorite, muscovite, and sericite. The various stages may represent a continuous sequence. The order, particularly in the case of the minor accessories and the latest quartz, may have been somewhat different.

INNER FACIES

COMPOSITION

The inner facies has generally been regarded as comprising the prevailing rock of the batholith; it is a calcic quartz monzonite with local variations to granodiorite and granite which, with few exceptions, merge into one another. Such local variations are ordinarily discernible only in thin sections but are locally distinguished by abundant flesh-colored potash feldspar and much muscovite, which signifies change to granite.

This facies exhibits less regional variation in composition than the marginal facies, being uniformly alkalic and silicic throughout the batholith. Toward the north end this facies is typically quartz monzonitic with some variations to granite (Anderson, 1930b, p. 21). Farther south in the Elk City region Shenon and Reed observed quartz monzonite with almost equal granodiorite (1934, p 16). Reed has found similar variations in the close-by Warren (1938, p. 7) and Florence (1939, p. 8) mining districts. Through other parts of the batholith the rock as at Edwardsburg (Shenon and Ross, 1936, p. 9), Yellow Pine (Currier, 1935, p. 8-10), Atlanta (Anderson, 1939, p. 7), Secesh (Capps, 1940, p. 7), and in the Casto quadrangle (Ross, 1934, p. 36) is mainly quartz monzonite, but granodiorite variations are locally almost as widespread as is the quartz monzonite. The facies is somewhat more calcic in the Boise Basin region. Much of it is granodiorite rather than calcic quartz monzonite (Anderson, 1942b).

ROCK DESCRIPTION

Most of the inner facies is much lighter-colored than the marginal having little or no evidence of gneissic structure which characterizes the marginal zone. Apparently the only exceptions so far observed are in the Yellow Pine (Currier, 1935, p. 8), Edwardsburg (Shenon and Ross, 1936, p. 9), and Casto (Ross, 1934, p. 36) areas where some slight banding has been pointed out. Much of the facies is typically light to medium gray, normally granitic but which may also be conspicuously porphyritic over broad areas, and in places somewhat aplitic. The light-colored grains commonly range from 4 to 7 millimeters in diameter, and the dark ones from 2 to 3 millimeters. The porphyritic rock differs from the normal only in being studded with Carlsbad-twinned potash feldspar phenocrysts as much as 30 millimeters long. In the aplitic variations the grains range from 2 to 3 millimeters. Because the aplitic variety forms such a small proportion it is probably ordinarily overlooked. The only mention of it has been by Currier at Yellow Pine (1935, p. 11) and by the writer in Boise Basin (Anderson, 1942a).

Unlike the marginal the inner facies contains little or no hornblende, sphene, and epidote, commonly less than 5 per cent biotite, and has oligoclase (Ab_{75} to Ab_{80}) as its characteristic plagioclase. It also has considerably more potash feldspar—orthoclase and microcline—commonly muscovite, and generally larger amounts of quartz than the marginal rock. Its accessories include allanite, zircon, apatite, magnetite, and in places a little garnet. Other minerals include chlorite and sericitic mica.

Analyses of typical quartz monzonite from various parts of central Idaho (Ross, *in* Currier, 1935, p. 8) show the following range of composition in percentages: quartz, 25 to 30; potash feldspar, 16 to 35; plagioclase, 20 to 45; biotite, 1 to 10. The typical rock ordinarily contains in percentages: 20 to 30 quartz, about 20 of potash feldspar, a little more than 40 of plagioclase, and a little less than 5 of biotite. These proportions agree fairly well with those obtained by the writer in the Boise Basin and other parts of the batholith, except that in some places the quartz exceeds 30 per cent of the rock, and muscovite is as abundant as the biotite. Muscovite is not distributed throughout the rock of the facies but in various localities may be inconspicuous or absent altogether. More sericitic mica as well as considerably more chlorite occur in this facies.

MINERAL RELATIONS

Mineral relationships in the inner facies are much the same as those of the marginal quartz diorite and granodiorite. Among the essential minerals oligoclase alone tends to be euhedral or subhedral, but it has regular borders only against other oligoclase grains. Otherwise its borders show all stages of penetration by potash feldspar, biotite, and quartz. It also is commonly retained as irregular residual grains and oriented inclusions in the larger crystals of orthoclase and microcline. Most of the oligoclase grains are somewhat dusted with small grains of sericitic mica or hold patches of sericitic mica in aggregates of somewhat larger grains. The oligoclase crystals in places hold widely scattered ghostlike inclusions of a zoned, somewhat more calcic plagioclase.

Most of the other minerals penetrate older minerals. The biotite, as in the marginal rock, tends to be interstitial between the plagioclase grains, but in general its outlines are just as ragged as those of the biotite in the somewhat older quartz diorite and granodiorite either because it is penetrated irregularly by andesine or because it itself has irregular protuberances in the andesine (Pl. 2, fig. 3). The orthoclase and microcline also penetrate and corrode the oligoclase as the microcline corrodes the andesine in the marginal facies (Pl. 2, fig. 3). Myrmekitic quartz in the oligoclase, corroded by potash feldspar, is even more abundant and striking. Optical characters indicate that locally the oligoclase in contact with the potash feldspar changed in composition. Potash feldspar is considerably more plentiful in this facies than in the marginal zone; its grains are appreciably larger here, so that it contains many more plagioclase inclusions. Many of these inclusions are preserved as widely scattered but uniformly oriented small islandlike remnants or as shadowlike ghosts (Pl. 2, figs. 3, 5).

Where the potash feldspar grains are least numerous, they merely embay or cut through the oligoclase (Pl. 2, fig. 3), but where the grains are the size of phenocrysts they may hold remnants of several oligoclase grains as well as scattered, corroded grains of quartz, as typical "groundmass" inclusions (Pl. 2, figs. 4, 5). Megascopically the potash feldspar phenocrysts appear to be subhedral to euhedral and to be bounded by crystal faces; actually, as viewed in thin section, their borders are minutely irregular and have small protuberances projecting unevenly into adjoining plagioclase grains. In some places the potash feldspar is orthoclase, in other places it is wholly microcline; but usually orthoclase and microcline occur together, commonly forming parts of a single grain.

Only a small part of the quartz is included in and corroded by the potash feldspar. The bulk of it occurs as large grains which irregularly vein the rock and send lobate tongues into and through the oligoclase and biotite and to lesser extent into the potash feldspar (Pl. 2, fig. 6; Pl. 3, figs. 1, 2, 3). As in the outer facies this quartz contains irregular inclusions of earlier quartz, and seams and lobes of it are embayed and cut by other seams and lobes (Pl. 3, fig. 3).

Large muscovite grains occur in either the oligoclase or the potash feldspar, commonly at the side of quartz seams and lobes. Its outline is not nearly so irregular as the biotite grains. It apparently replaced the feldspars as euhedral grains.

Such minor accessories as magnetite, zircon, apatite, garnet, and allanite are mostly distributed along borders and cleavages of the feldspars and biotite and along margins of the quartz seams and lobes. Some of the chlorite forms entire or partial pseudomorphs after biotite. The sericitic mica occurs as scattered patches in the oligoclase and in lesser amounts in the potash feldspar. In places it appears as thin seams extending into and even across the quartz grains and seams.

INTERPRETATION OF MINERAL RELATIONS

Since the mineral relationships in the inner facies are practically identical to those in the marginal rock they are similarly interpreted. Were the oligoclase grains not so thoroughly embayed, corroded, and engulfed by the potash feldspar, they would possess partial crystal outline or one determined solely by crowding or mutual interference during crystal growth. Likewise, the biotite between and around the oligoclase grains suggests rather early crystallization, but its irregular penetrating borders argue also for a later growth. Wedge-shaped grains of orthoclase and quartz between some of the oligoclase crystals, or with the oligoclase as remnants within larger grains of corroding microcline, also suggest early crystallization. The following minerals therefore suggest normal, orderly crystallization from a consolidating magma: biotite, oligoclase, orthoclase, and quartz.

Another assemblage of minerals may be interpreted as replacements in the crystallized rock. The younger generation of minerals includes minor amounts of biotite, much orthoclase and microcline, two generations of quartz, and accessory apatite, zircon, and magnetite, in places allanite and garnet, and finally a little chlorite, and considerable muscovite and sericite. As the microcline commonly merges with orthoclase, probably the orthoclase formed earlier and at a higher temperature; both were deposited by replacement in a succession determined by temperature, the microcline perhaps by partial inversion of the orthoclase.

APLITE

DISTRIBUTION AND STRUCTURAL RELATIONS

Most of the aplite dikes are concentrated in relatively small areas in and adjacent to some of the mining districts within the batholith. The dikes are fairly numerous in the Gambrinus district in Boise Basin near Idaho City (Anderson, 1942a) and are exceptionally abundant in the Atlanta district (Anderson, 1939, p. 8). Numerous dikes are also reported near Yellow Pine (Currier, 1935, p. 7, 11-12), but in general aplites are scarce. Their distribution appears to be structurally controlled by certain zones of weakness developed during closing stages of consolidation of the batholith.

The aplitic bodies are ordinarily small,—usually several inches thick, exceptional ones in Boise Basin range up to 200 feet in thickness. The aplites in general are intrusions guided by fissures and ordinarily are tabular and have steep dips. Contacts between the dikes and the bordering walls are sharp.

COMPOSITION

As most of the aplites contain about equal potash feldspar and plagioclase they are quartz monzonite aplites. A few containing considerably more plagioclase than potash feldspar are granodiorite aplites. Whether any of the bodies contain a preponderance of potash feldspar and should be classed as granite aplites is a problem for future investigation. In some bodies grains of potash feldspar are larger and more abundant and express textural transition into pegmatites.

ROCK DESCRIPTION

The aplite is lighter-colored and considerably finer-grained than the prevailing rock of the batholith. In some of the larger dikes its average grains are about 2 millimeters in diameter, but in the smaller dikes grains average only about 1 millimeter or less. The unweathered rock is white, or pinkish, but the weathered somewhat resembles fine-grained brownish sandstone.

Most of the aplites apparently have about equal amounts of quartz, sodic oligoclase (Ab_{85} to Ab_{90}), and microcline and orthoclase, and less than 5 per cent of biotite, most of which is bleached or changed to muscovite. Some aplites contain sphene and also variable amounts of accessory garnet, zircon, apatite, allanite, and magnetite, and a little sericite, chlorite, and calcite. It differs essentially from the normal quartz monzonite and granodiorite in its finer grain, its occurrence, and in its more sodic plagioclase.

MINERAL RELATIONS

Although the texture is typically aplitic, the mineral relationships resemble those of the main rock of the batholith. Except for embayed margins, the plagioclase forms subhedral or euhedral crystals, and the biotite grains, if any protuberances were eliminated, would be largely subhedral. Most of the plagioclase grains are embayed by small lobes and seams of quartz, are penetrated and corroded by grains of potash feldspar, and are commonly penetrated by small slivers of biotite. The small amount of potash feldspar not invading other minerals forms small wedge-shaped grains between the plagioclase crystals. Some of the quartz does not embay or cut the plagioclase but forms wedge-shaped grains, many of which have later been engulfed in grains of microcline. Most of the accessory minerals are aligned along grain contacts and mineral cleavages. Some biotite is penetrated by or is replaced by chlorite. The biotite, chlorite, and feldspar are penetrated or replaced by muscovite and sericite. Some of the sericite is scattered as granular aggregates in patches through the plagioclase. The calcite is in seams cutting the rock or in scattered grains.

INTERPRETATION OF MINERAL RELATIONS

Such small differences as may exist between the aplite and the normal rock are largely textural because of the finer grain and the presence of a somewhat smaller proportion of minerals that tend to penetrate other minerals. If the effects created by mineral penetration were eliminated, the rock would show euhedral and subhedral plagioclase, subhedral biotite, and anhedral orthoclase and quartz in the interstices. Thus the relations of minerals would be those in normally consolidated rock. These minerals, however, particularly the plagioclase, have been slightly penetrated by fringes of biotite, embayed and corroded by grains of potash feldspar, and cut and indented by lobes and seams of quartz. Also, the earlier grains have been penetrated along contacts and cleavages by small crystals of zircon, apatite, magnetite, and garnet, and some of them have been partly changed to chlorite, muscovite, sericite, and calcite. These changes must have been made by replacement of the minerals penetrated.

The aplites thus are the product of both crystallization and later replacements. After crystallization of plagioclase, biotite, and some potash feldspar and quartz another succession was added by replacement, a succession that included a little biotite, potash feldspar, quartz, and the various accessory and the so-called secondary minerals. Some biotite was added first, closely followed by potash feldspar and quartz. The different accessories were apparently deposited during the stage of quartz deposition, but, because the crystals were small and not in contact with each other, their precise order could not be determined. The chlorite, muscovite, sericite, and calcite sequence is more easily established and apparently marks the end of the replacements.

PEGMATITE

DISTRIBUTION AND STRUCTURAL RELATIONS

Pegmatitic seams and dikes are more widespread than the aplitic dikes but are more common at the roof and marginal zones of the batholith. Some of the pegmatites are related genetically to the marginal rock and in places are abundant in the outer facies. Others are related to the inner rock occurring at the outer margin of the inner facies and in the bordering marginal facies. The two series of pegmatites overlap in their distribution in the marginal zone and are not always easily segregated. Along the northwest margin of the batholith the compositional contrast is clearly marked so that they may be separated, but in the southwestern part where no such contrast exists the two cannot be satisfactorily distinguished.

Where the pegmatites appear in the aplitic zones, as in the Atlanta district, they cut the aplite dikes or lie in and along them as though the fissures that guided the aplitic magma reopened at the time of pegmatite formation. Because of such reopening many of the aplites appear to grade into pegmatites. The pegmatites merge with or grade into the surrounding granitic and aplitic country rock.

COMPOSITION

Most of the pegmatites have about the same composition as the genetically related granitic rocks, and are designated as quartz monzonite pegmatite, granodiorite pegmatite, and quartz diorite pegmatite. Those related to the quartz diorite along the northwest margin of the batholith have been described as hornblende-quartz diorite pegmatite and biotite-quartz diorite pegmatite, and those associated with the inner quartz monzonite are quartz monzonite pegmatite and granite pegmatite (Anderson, 1930b, p. 19, 22). Most of the pegmatites associated with the marginal rock of the southwest range from granodiorite to quartz monzonite. Pegmatites related to the inner rock range from quartz monzonite to granite. In general the pegmatites tend to be alkalic, and several are considerably more alkalic than the parent facies and than other pegmatites in their own vicinity. Because the pegmatites around the northwest margin of the batholith have not been studied in as much detail as those in the southwest part they will not be discussed further; the following discussion deals with those examined in southwestern Idaho.

ROCK DESCRIPTION

Although graphic intergrowths of quartz and feldspar are not uncommon, most of the pegmatites resemble coarse-grained granite with the grains ranging from less than 1 inch to 6 inches in diameter. The dark minerals are invariably less abundant than in the related quartz monzonite and granodiorite, and as a result the rock is lighter. Some dikes are composed almost wholly of coarse crystals of pink or white feldspar and are uniformly white or pink, but in most dikes the rock is mottled because of many grains of smoky quartz and widely scattered crystals of black biotite.

The most abundant minerals are quartz, microcline, plagioclase, and small but variable amounts of biotite and muscovite. Most of the dikes also contain scant amounts of apatite, zircon, magnetite, pyrite, locally garnet, and chlorite and sericite. In pegmatites associated with the marginal facies, the plagioclase is andesine; in other pegmatites, generally oligoclase, but in some it is albite. The proportions of the essential minerals are variable. Quartz is relatively abundant in most of the dikes; except in the pegmatites related to the marginal facies, the microcline is usually more plentiful than the oligoclase. Many of the highly feldspathic dikes are composed almost entirely of microcline. Widely scattered crystals of biotite and muscovite, ranging up to 3 inches in diameter, are more common in the more siliceous pegmatites. Other minerals, except locally small crystals of garnet, are as a rule very inconspicuous.

MINERAL RELATIONS

Most of the plagioclase and some of the grains of quartz are completely enclosed in the microcline (Pl. 3, fig. 4),—some of the enclosed grains altered to shadowlike inclusions. These inclusions commonly match the plagioclase and quartz grains in the adjacent granite and aplite and actually merge with them at the contact, the grains of microcline penetrating and isolating remnants of the invaded rock. The plagioclase grains are also penetrated by crystals of biotite which may even project into the wall rock. The biotite, however, is not penetrated by the microcline. Albite forms perthitic spindles in the microcline or crystals which project into the microcline grains. Although some of the quartz occurs as residual grains in the microcline, most of it is in grains and seams that penetrate the biotite, plagioclase, and microcline. It is lobate like much of the quartz in the marginal and inner facies. Some of the quartz cuts the large microcline grains in irregular veinlets that locally simulate graphic intergrowths. The muscovite and most of the other minerals lie almost wholly within the feldspar grains.

INTERPRETATION OF MINERAL RELATIONS

With the exception of the inclusions of plagioclase and quartz that obviously were derived from the older rocks, the writer believes all the minerals were formed by replacement. The biotite that projects through the residual masses of included wall rock, the microcline that retains numerous residual inclusions of the country rock, the albite that penetrates the microcline, and the lobate quartz that cuts and embays them all, permit no other interpretation. Deposition of the quartz apparently persisted until after the muscovite and all the accessory minerals had been added.

ENDOMORPHISM

END-STAGE PHENOMENA

Mineral relations of the marginal and inner facies and of the associated aplite and pegmatite dikes show that the present rock has resulted partly from replacement. Each facies and associated aplites, except the pegmatites, have had much the same minerals added. The pegmatites differ because the replacements have not been superposed upon crystallized minerals of a pegmatitic magma but have acted upon the minerals of the inner and marginal facies and of the aplite.

Such changes have been cited by Gillson (1927, p. 5-8, 16-30), Currier (1935, p. 8-11), and Anderson and Hammerand (1940, p. 585-586) as end-stage replacements, taking place during the closing stages of consolidation. Considerable but variable amounts of potash and silica and minor amounts of other materials have been contributed, which must have come from the batholith as emanations from the deep magma.

ENDOMORPHISM BY POST-CONSOLIDATION SOLUTIONS

Gillson (1927, p. 16-30) and Currier (1935, p. 8-9), who have described the changes in parts of the batholith regard some of the changes as deuteric forming by reaction of the crystals with residual liquid. Gillson then has described subsequent changes as pneumatolytic and hydrothermal, and Currier, as wholly hydrothermal. It appears to the writer that the late assemblage of biotite, potash feldspar, quartz, sphene, zircon, apatite, magnetite, allanite, epidote, garnet, chlorite, muscovite, and sericite is typically hydrothermal and accords with a gradual decline in temperature. Possibly the reaction rims on some of the plagioclase and the myrmekitic blebs of quartz in the plagioclase grains may have been produced by the reaction of crystals with residual liquid (Currier, 1935, p. 8-9); however, since these deuteric effects are so closely associated with replacement, they may have been produced during the succeeding post-consolidation stage. In view of the considerable amounts of material added to the rock, the changes must have taken place primarily under the influence of hydrothermal solutions. These solutions were probably residual liquids in the magma at a somewhat greater depth, perhaps at much greater depth in the case of the solutions which produced the pegmatites. Because of the difficulty of distinguishing between alteration of deuteric origin and replacement during an immediately succeeding high-temperature hydrothermal stage, and since any relation that might be construed as deuteric constitutes such a minor part of the magmatic after-effects, no distinction between the deuteric and the hydrothermal is made in this paper. The stages of alteration may have been represented by a continuous sequence conforming to a gradual change in temperature and chemical character of the hyperfusible constituents which became concentrated during the later stages of consolidation and then emanated from depth while the cooling, consolidation, and fracturing of the batholith proceeded downward.

ENDOMORPHIC PROCESSES

Such changes as have taken place in the consolidated, or largely consolidated, rocks might be described as the product of end-stage metamorphism or post-consolidation metamorphism. However, the term endomorphism accords more with common usage. The main process of endomorphism has been silicification as it has been in the Sparta granite (Gilluly, 1933, p. 65-81). Large amounts of silica have been added to the rock of the Idaho batholith, and there has also been a variable enrichment in potash. The addition of potash feldspar has caused compositional variations within both facies and in the aplite and pegmatite.

RELATION OF ENDOMORPHISM AND EXOMORPHISM

Where emanations have invaded country rock, changes quite analogous to those within the intrusive rock have been produced. If the invaded rocks were quartzitic they have become rather extensively granitized. Such contact effects have been described around the margin of the Cassia batholith (Anderson, 1934a, p. 376-392) and were a chief factor in the development of the batholith in the Selkirk Mountains in northern Idaho where emanations have changed the Belt sedimentary rocks almost entirely into a distinctly banded gneiss ranging from granite to quartz diorite (Anderson, 1940, p. 17-20). Similar contact phenomena also have been described for the Idaho batholith in north Idaho (Anderson and Hammerand, 1940, p. 561-589). The problems of endomorphism are probably closely related to those of the contact metamorphism and granitization.

CONSOLIDATION HISTORY

The consolidation history of the Idaho batholith has been complicated by the two-stage emplacement. In spite of the solidification of the earlier marginal rock under somewhat greater stress than the later inner facies and despite the fact that a time interval of unknown duration permitted an appreciable change in the composition of the inner mass before its emplacement and consolidation, the endomorphism of the rock of each facies has been remarkably similar. Evidence indicates that the original marginal facies was a diorite with only minor amounts of quartz but locally with hornblende or biotite, or both, and with andesine. As a result of end-stage contributions, some biotite was added, as were small but variable amounts of potash feldspar, much quartz, generally exceptionally large amounts of sphene, and such minor minerals as apatite, zircon, magnetite, allanite, epidote, chlorite, locally muscovite, and sericite. The most important contribution was silica, and the rock was in general thoroughly silicified. Where little or no potash feldspar was added, as along the northwest margin of the batholith, a quartz-rich diorite (tonalite) was produced, but where considerable amounts of potash feldspar were added, as along the south and southwest margin, granodiorite formed.

The consolidation history of the inner facies differs but slightly in detail from that of the marginal. Its magma was apparently somewhat less calcic than that of the marginal rock and yielded an original rock ranging apparently from diorite or quartz-bearing diorite to granodiorite with oligoclase, rather than andesine. Post-consolidation emanations added considerable silica and potash, increased slightly the amount of biotite, and introduced the materials for the formation of the accessories. More potash feldspar (orthoclase and microcline) was added than in the marginal rock, thus changing much of the original rock to quartz monzonite. Where less potash feldspar was added, the conversion was to granodiorite. Locally enough potash was added to form granite, in places a muscovite granite. Over wide areas the added feldspar formed replacement phenocrysts that make the rock conspicuously porphyritic. As in the marginal rock silicification has been the dominant process in the endomorphism, but the process by which potash feldspar was added also has been important.

Consolidation of the aplite dikes has followed a course similar to that of the main granitic facies. Progressive cooling of the batholith inward and downward apparently was accompanied by shrinking and cracking of the solidified roof zone. When avenues were opened to greater depth by deeper-seated fissuring, the magma, which had become less calcic, perhaps because of longer-continued differentiation, was injected upward into the fractures near the top, where it consolidated as aplite. Then the aplite was considerably endomorphosed by associated emanations and was made somewhat more silicic and alkalic.

Pegmatites, on the other hand, apparently did not originate directly from consolidating magma as did the aplites and granitic facies but appear to be wholly hydrothermal and associated with replacement of the earlier consolidated granitic and aplitic rock, by the action of silica and potash-rich fluids from deeper sources than those that endomorphosed the aplites.

CONCLUSIONS

That the composition of intrusive igneous rocks may be considerably modified by the action of end-stage emanations is rather generally recognized, but the extent to which such action might affect huge batholithic bodies has probably not been fully appreciated. The endomorphism that accompanied the consolidation of the Idaho batholith may be more than locally significant and may be a feature associated prominently with batholiths in other localities.

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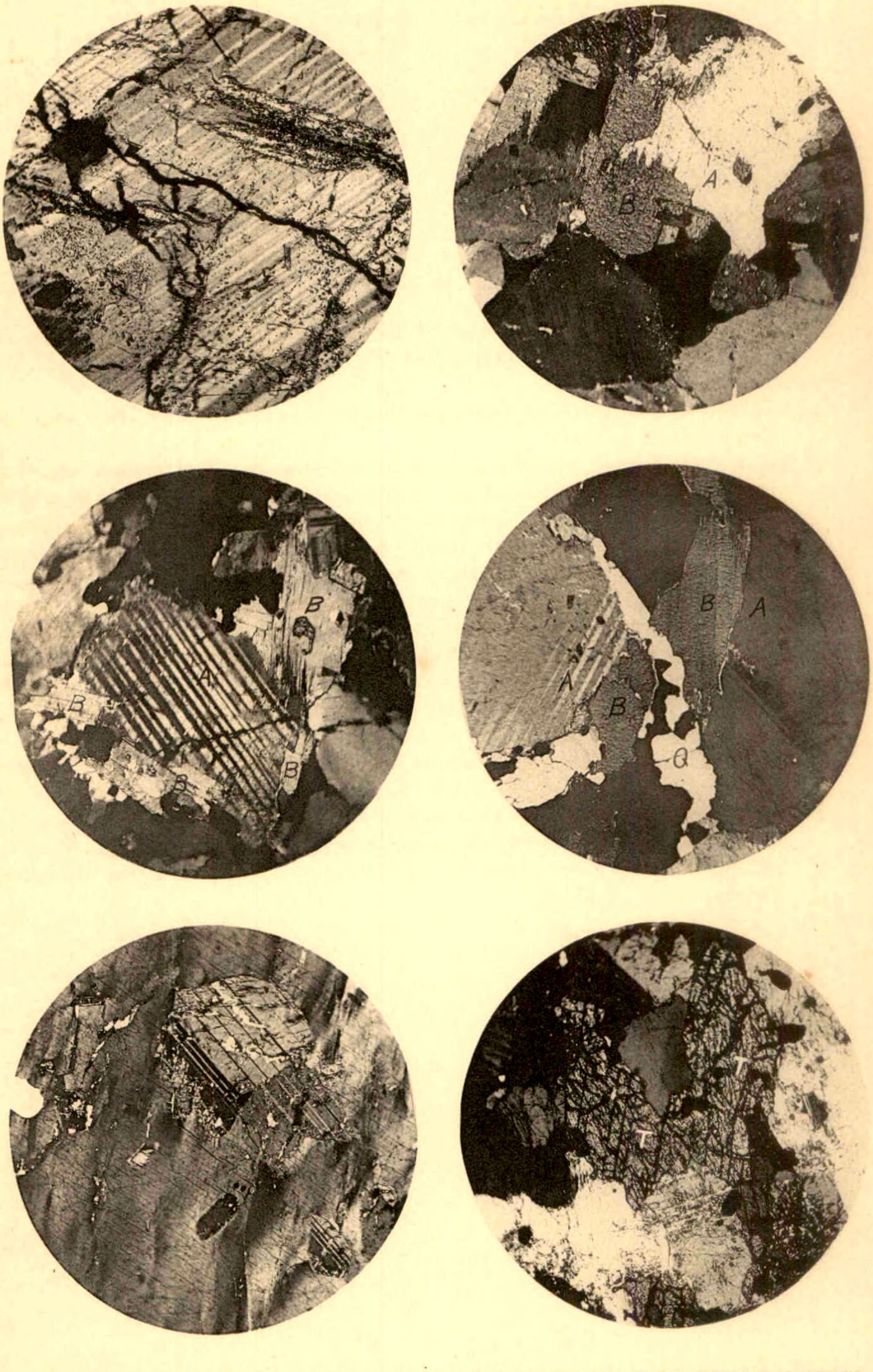
PLATE 1

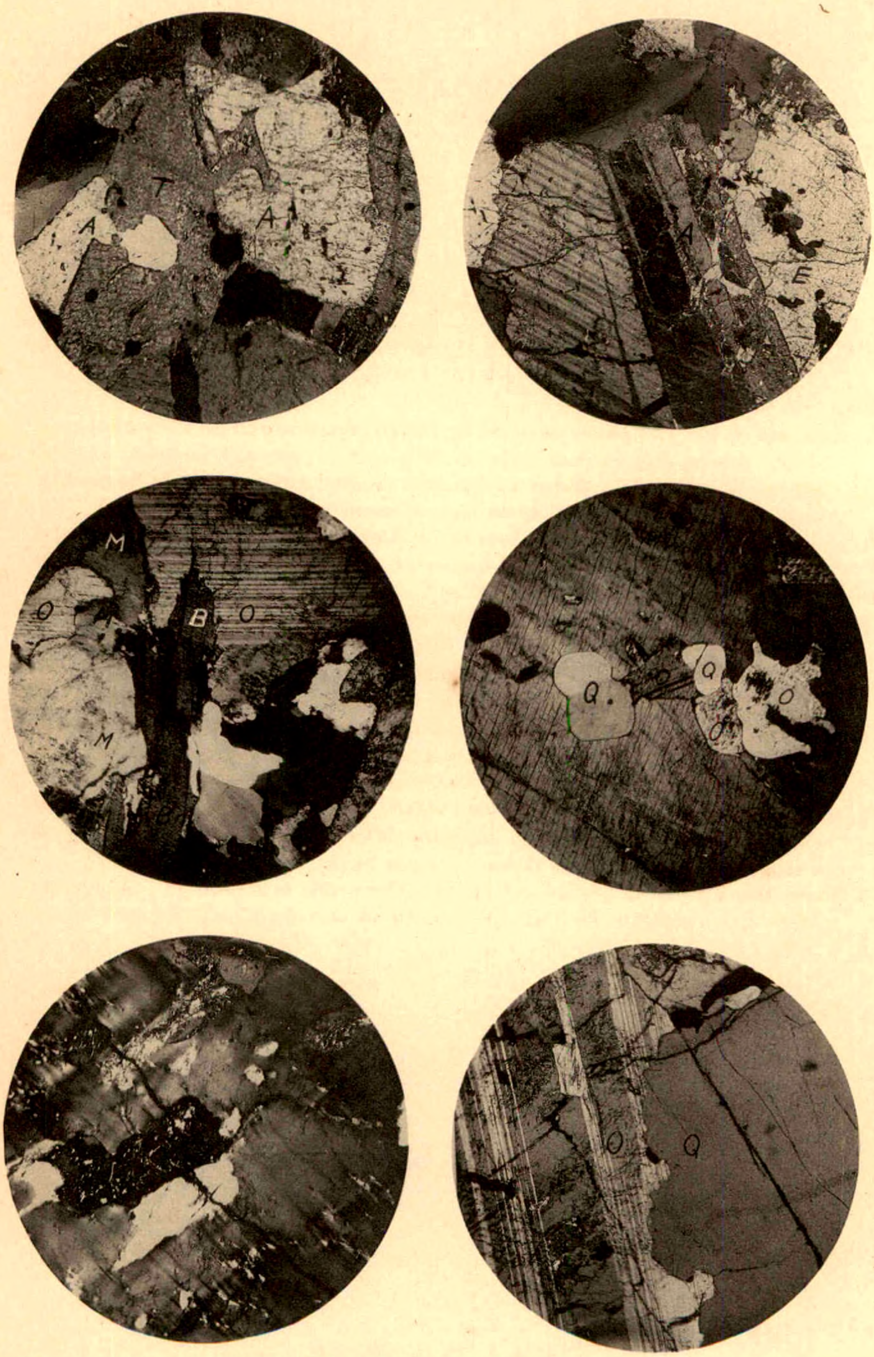
PHOTOMICROGRAPHS OF MINERAL REPLACEMENTS IN MARGINAL FACIES

Figure

1. Andesine crystal containing irregular remnant of an earlier, more calcic andesine Crossed nicols, × 28
2. Biotite (B) showing irregular penetration by andesine (A) Oriented islands of the biotite remain as inclusions. This biotite is interpreted as early and as partly replaced by the andesine. Generally, contact between invading andesine and biotite is more regular. Biotite may have been inherited from "granitized" quartzite. Crossed nicols, × 28.
3. Andesine crystal (A) irregularly penetrated at its margin by crystals of biotite (B). Biotite crystal in northeast quadrant also penetrates and partly replaces a second andesine grain. The biotite is interpreted as a late addition to the rock during a post-consolidation stage of alteration. Crossed nicols, × 37
4. Biotite (B) between andesine grains (A) cut and partly replaced by a vein of late post-consolidation quartz. Crossed nicols, × 28
5. Microcline which contains remnants of partly replaced andesine grains Crossed nicols, × 28
6. Sphene crystals grouped between and along feldspar grains. The sphene assumes its own crystal form in replacing other minerals. Replacement interpreted as taking place during an early stage of post-consolidation alteration Crossed nicols, × 37.



PHOTOMICROGRAPHS OF MINERAL REPLACEMENTS IN MARGINAL FACIES



PHOTOMICROGRAPHS OF MINERAL REPLACEMENTS IN MARGINAL AND INNER FACIES

PLATE 2

PHOTOMICROGRAPHS OF MINERAL REPLACEMENTS IN MARGINAL AND INNER FACIES

Figure

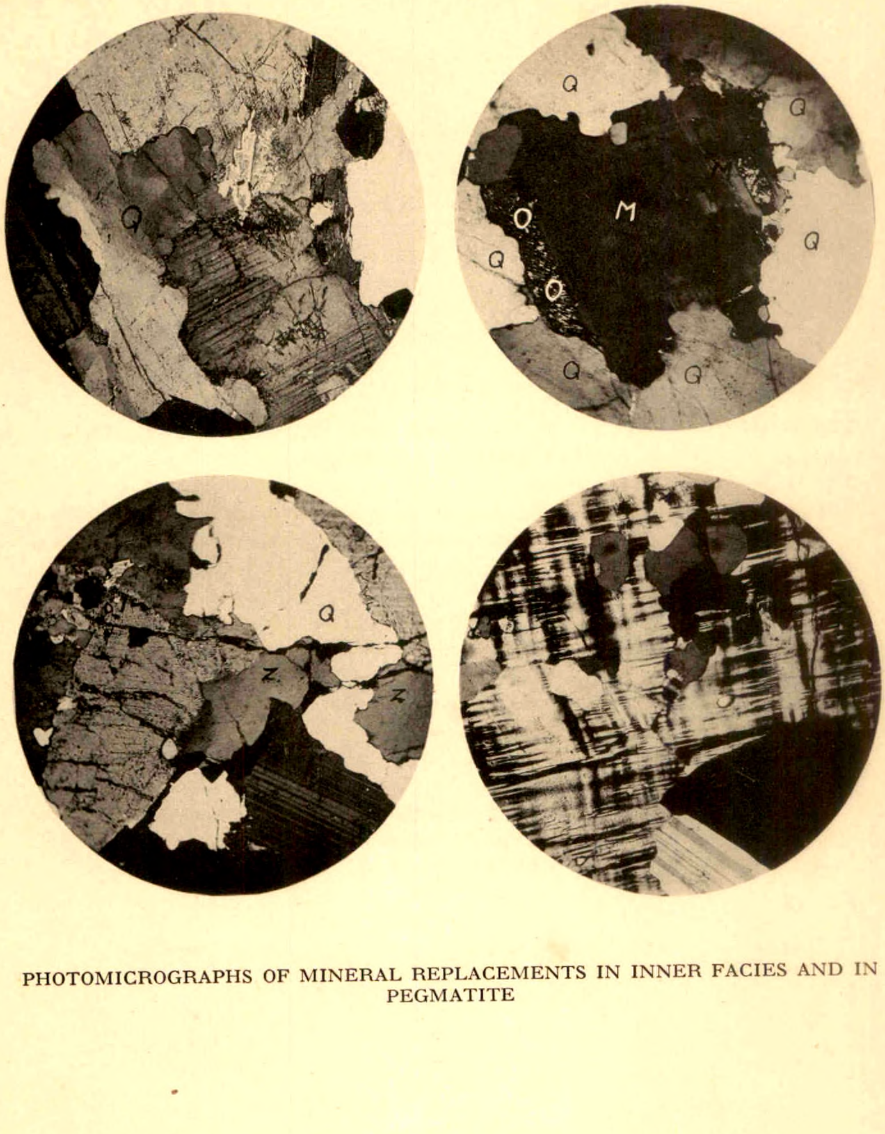
1. Andesine crystal (A) partly included in a large crystal of sphene (T). Apophysis of the sphene also extends outward into another grain of andesine (right). Sphene interpreted as added to the rock during an immediately succeeding stage of post-consolidation alteration. Crossed nicols, × 28.
2. Large grain of epidote (E) with apophyses extending into and replacing a crystal of andesine (A). Affords an example of post-consolidation replacement in marginal facies. Crossed nicols, × 28.
3. Biotite (B) projecting into and partly replacing oligoclase crystal (0) and in turn partly embayed by microcline. Vein of microcline (M) also cuts entirely through the oligoclase crystal, isolating a part of it from the main crystal. Example of replacement phenomena in the inner facies. Crossed nicols, × 37.
4. Large grain of orthoclase with remnant inclusions of earlier oligoclase (0) and quartz (Q). Orthoclase is interpreted as an addition to the rock of inner facies during period of end-stage alteration. Crossed nicols, × 37.
5. Large microcline crystal with remnant quartz grains (light) and shadowlike inclusions of original plagioclase feldspar. Microcline is interpreted in part as a replacement of the inner facies. Crossed nicols, × 37.
6. Shows the way the late-stage quartz (Q) embays the oligoclase (0). Replacing lobes may penetrate far into the interior of the oligoclase. Crossed nicols, × 28.

Plate 3

PHOTOMICROGRAPHS OF MINERAL REPLACEMENTS IN INNER FACIES AND IN PEGMATITE

Figure

1. Seam of lobate quartz (Q) penetrating in and between crystals of oligoclase. Uneven borders of quartz grains suggest replacement rather than interstitial filling. Typical of much of the late quartz of the batholith. Crossed nicols, × 28.
2. Microcline grain (M) penetrated unevenly on three sides by lobate quartz (Q) and protected on fourth side by similarly penetrated grain of oligoclase (O). Crossed nicols, × 37.
3. Seam of early quartz (Z) cut by seam of somewhat younger quartz (Q). Both penetrate and replace oligoclase grains. The younger quartz (Q) also embays and replaces the older quartz (Z). Two late stages of quartz deposition is suggested. Crossed nicols, × 28.
4. Microcline grain in pegmatite containing remnant inclusions of quartz and oligoclase inherited from the quartz monzonite. Microcline is interpreted as replacing the wall rock. Crossed nicols, × 28.



PHOTOMICROGRAPHS OF MINERAL REPLACEMENTS IN INNER FACIES AND IN PEGMATITE

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PLEISTOCENE GEOLOGY OF WESTERN CAPE COD, MASSACHUSETTS

BY KIRTLEY F. MATHER, RICHARD P. GOLDTHWAIT, AND LINCOLN R. THIESMEYER

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ABSTRACT

During recession of continental ice of the Wisconsin stage a complex series of moraines and pitted plains was deposited on Cape Cod and adjacent portions of southeastern Massachusetts Most notable are the Buzzards Bay moraine, extending in a northeast-southwest direction along the southeast shore of Buzzards Bay, and the Sandwich moraine, trending east-west and roughly parallel to the south shore of Cape Cod Bay. These were deposited during minor readvances of the lobate margin of the ice The lithologic character of the transported stones reveals the direction from which the ice advanced and is distinctive for the two major lobes of ice. Along the shore north of the Sandwich moraine there are discontinuous bodies of till resting upon and surrounded by glacio-fluvial deposits. These constitute the Scorton moraine, deposited during the wastage of essentially stagnant ice, after retreat of the ice-front from the position marked by the Sandwich moraine. The Ellisville moraine, north of Buzzards Bay, represents a third type of recessional moraine, resulting from a temporary halt in the retreat of the margin of an active ice-lobe without readvance.

The Mashpee pitted plain occupies the angle between the Buzzards Bay and Sandwich moraines. It consists largely of sand and gravel spread by melt water across a land surface strewn with many large blocks of stagnant ice, left stranded beyond the continuous, lobate margin of the ice sheet, at a time when that margin was an appreciable distance north and northwest of the line now marked by those moraines and prior to the readvance of the ice responsible for their emplacement. An extraordinary number of large and deep kettle holes dot the surface of the gently sloping fan-shaped plain. Adjacent to some of them, clumps of till are embedded in the glacio-fluvial deposits Many of the buried blocks of ice responsible for the kettle holes did not melt away until after the surface of the plain had been extensively furrowed by subaerial erosion and its western and northern margins had been overridden by active ice readvancing to form the bordering moraines

Thousands of ventifacts are embedded in undisturbed glacial and glacio-fluvial deposits and in disturbed surficial zones throughout the region. All types of wind abrasion from incipient polishing and faceting to pronounced etching, deep fluting, or pitting are displayed. These stones were sculptured by sandblasting in a periglacial climate, both prior to the deposition of the till and during the construction of the glacio-fluvial deposits. Many of the wind-scoured surfaces were modified to greater or lesser extent by abrasion during glacial or glacio-fluvial transport. Thin weathered rims on some of the ventifacts apparently survived such transport but were truncated by wind-cut facets

INTRODUCTION

WORK ON WHICH THIS REPORT IS BASED

Field investigations in the western part of Cape Cod were begun in June 1939 under the auspices of the United States Geological Survey as a part of a co-operative project for geological research in Massachusetts, financed jointly by the Commonwealth and the Federal Government. Recently completed topographical maps with a scale of 2 inches to the mile were used as the base for the detailed geologic mapping which was extended over most of the area covered by the Sagamore, Sandwich, Pocasset, Falmouth, and Woods Hole quadrangles and the eastern part of the Onset quadrangle during the summer of 1939.

ACKNOWLEDGMENTS

During the progress of the field work and preparation of this report, the writers were greatly aided by numerous conferences in the field and office with L. W. Currier, geologist in charge of the co-operative project, Robert W. Sayles, and Kirk Bryan. Arthur S. Knox studied many localities and is responsible for the identification of the animal and plant remains which give a clue to the fauna and flora of the region during late Pleistocene and Recent time. Edward A. Schmitz prepared the maps and diagrams for publication. For the photographs of ventifacts taken in the laboratory the writers are indebted to Edward S. Wood, Jr., of the Institute of Geographical Exploration at Harvard University.

EARLIER WORK ON CAPE COD

The glacial, glacio-fluvial, and glacio-lacustrine deposits comprising nearly all the land mass known as Cape Cod have long been of great interest to geologists and have been the subject of many reports published during the last hundred years. The abundant erratics strewn over the surface of the Cape and the adjacent islands toward the south and southwest were first described in geological literature by Hitchcock in 1833 (p. 143-144) as a part of the "diluvium" resulting from "the flood," although he evidently had grave doubts about the adequacy of that explanation of their origin. In 1879, Upham (1879, p. 81-92, 197-209) described many of the features discussed in this paper as a part of the more northerly of two glacial moraines that he had traced eastward and northeastward from Long Island. Commenting upon Upham's work, T. C. Chamberlin (1883, p. 377-379) stressed the lobate character of the ice sheet from which these moraines had been deposited and inferred that the moraines of Cape Cod marked a recessional stage of the ice that had retreated not long before from its terminal position on Martha's Vineyard and Nantucket.

Shaler (1881, p. 57-58; 1889, p. 46; 1898, p. 497-593) likewise recognized the glacial character of the Cape Cod landscape and described many of the features in the western part of Cape Cod as portions of a frontal moraine and its marginal outwash apron. In 1906, Wilson (p. 4, 7, 48-50) presented an argument for the "probable existence of a great Newfoundland ice-sheet which was the source of the ice on Nantucket and on Cape Cod" and reported an "interlobate moraine extending in a southerly direction from the neighborhood of Plymouth and Manomet Hill to Wood's Hole." He described the portion of the "Cape Cod moraine" parallel to Buzzards Bay as partly interlobate and partly marginal to the "Long Island lobe" of ice at its second stage following retreat from the northwest side of Martha's Vineyard.

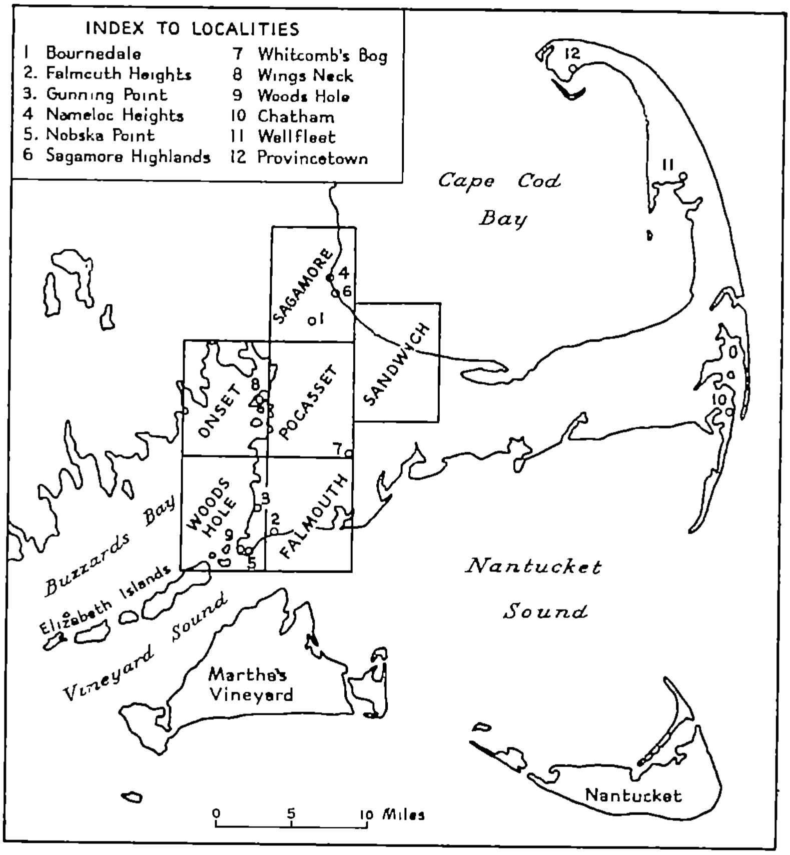


FIGURE 1.—*Outline map of southeastern Massachusetts*

Including Cape Cod. Quadrangles studied in detail are indicated by name

During the same decade, Woodworth, Fuller, Clapp, and others were investigating the glacial features of Long Island, Rhode Island, and Massachusetts, and frequent references were made to the possible relationships between the moraines of Cape Cod and those of other localities.

In 1925, Johnson (p. 105-117) presented a convenient synthesis of the ideas thus gained and identified the moraines with which the present paper deals as a part of the "Harbor Hill moraine" extending eastward from Orient Point on Long Island. Although Woodworth's field studies were completed in 1919, his results were not published until 1934, in a memoir that embodies the most detailed and inclusive report on the geology of Cape Cod thus far available. In this report all the morainic deposits south and southeast of the Cape Cod canal in the area now under consideration are grouped together as the "Falmouth moraine," formed during late Wisconsin time after the ice margin had retreated northward from the outer moraine on Martha's Vineyard and Nantucket.

More recently, Bryan (1932, p. 176) directed attention to the "weathering, wind-worn stones and intense frost heaving . . . characteristic of many localities of southeastern Massachusetts" and suggested that "Martha's Vineyard, parts of Cape Cod and of the mainland of Massachusetts were, on this evidence, free of ice in the late Wisconsin and perhaps even through the whole of Wisconsin time." Moreover, Sayles (1939, p. 1931-1932), working principally on the more easterly portion of the Cape, observed phenomena indicative of an even greater complexity of glacial and associated deposits than had been inferred previously. It therefore seemed desirable to take advantage of the recent completion of detailed topographic surveys of the region adjacent to Buzzards Bay and to re-examine this critical area in the light of modern concepts of Pleistocene glacial geology.

PART I. GLACIAL MORAINES IN THE VICINITY OF BUZZARDS BAY

BUZZARDS BAY MORAINE

As indicated in Figure 1, the quadrangles surveyed during the 1939 field season cover the broad base of Cape Cod, nearly separated from the mainland of Massachusetts by Buzzards Bay, and the "spur" extending southwestward toward the Elizabeth Islands. The most conspicuous topographic features are the hills and dales, knobs and hollows of the boulder-strewn belts designated in this paper (Fig. 2) as the Buzzards Bay moraine and the Sandwich moraine. These moraines are connected but diverging portions of the extensive compound, morainal belt stretching from the westernmost of the Elizabeth Islands to the eastern shore of the "fore-arm" of Cape Cod, described by Woodworth as the "Falmouth moraine." The Buzzards Bay moraine was formed by the Buzzards Bay lobe of the Labrador ice sheet; the Sandwich moraine by the Cape Cod Bay lobe of the same great glacier.

The Buzzards Bay moraine extends in almost a straight line from the north-central part of the Pocasset quadrangle southward and

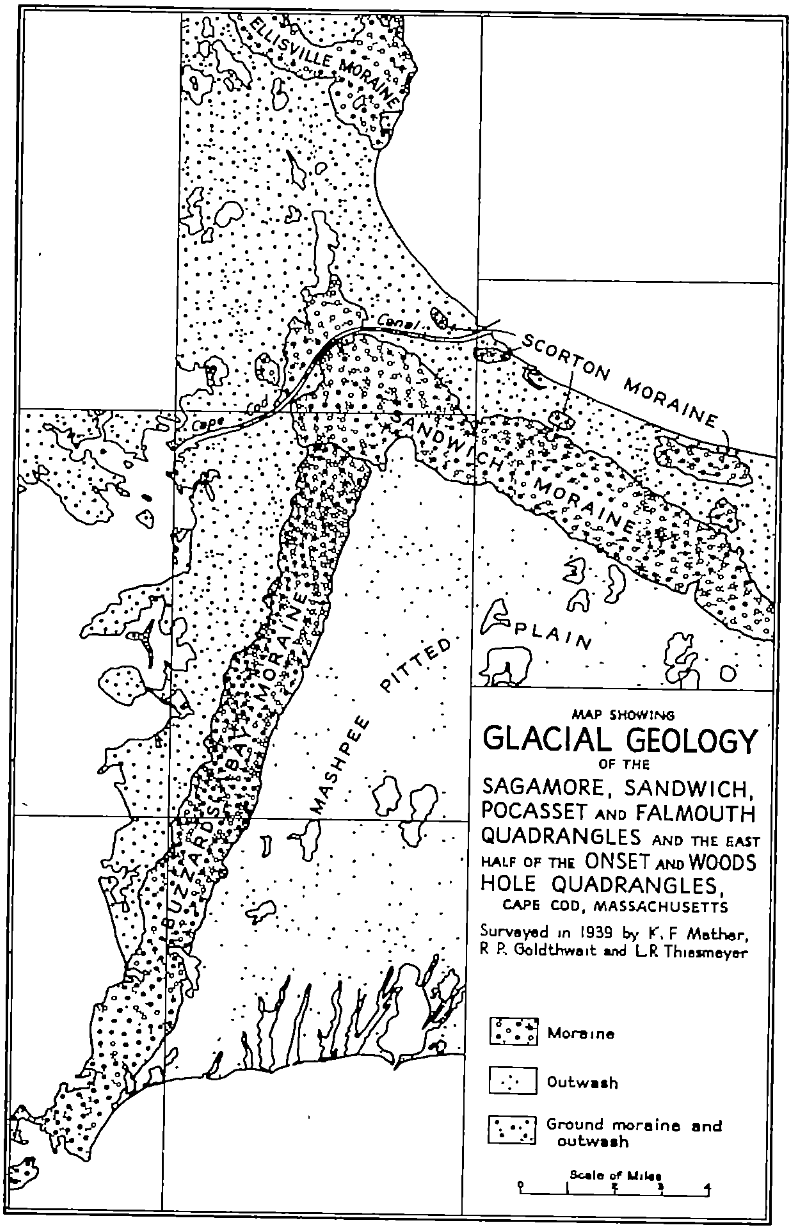


FIGURE 2.—*Map showing glacial geology of western Cape Cod*

southwestward across the northwest corner of the Falmouth quadrangle to Woods Hole, near the southeast corner of the Woods Hole quadrangle. Thence it presumably curves westward in the Elizabeth Islands, which the writers have not as yet studied. Its eastern margin is marked nearly everywhere by an abrupt change in topography and lithology, from the rough, hilly landscape of the moraine, liberally strewn with large erratic blocks and boulders, to the flat, gently sloping surface of the Mashpee pitted plain with its many deep kettle holes and its cobblestones, gravel, and sand. The western margin is much less definite; here there are gentler and more irregular slopes from the higher ground of the moraine to a gravel-strewn area with an occasional patch of till, leading down to the shore of the bay.

An observer driving southward on the highway connecting Buzzards Bay and Falmouth (State Highway 28, Fig. 3), from a point about one mile south of Bourne Bridge, sees the knobby, ridgelike moraine forming the skyline toward the east. At first the margin of the moraine is half a mile from the highway, but the two converge and, for several miles southward from the crossing of the Pocasset-Forestdale road, the highway follows rather closely the morainic boundary. South of West Falmouth, however, the highway crosses obliquely through the moraine to Falmouth, just beyond the eastern front of the moraine on the Mashpee pitted plain. Similarly the Buzzards Bay shore line and the moraine converge southward, so that in the Saconesset Hills of east-central Woods Hole quadrangle the moraine rises directly from the narrow sandy beach and continues thence to Woods Hole along the shore.

Many of the irregular hills in the northern part of the moraine rise to altitudes of 250 to 300 feet; the highest, Signal Hill, is 306 feet. The bottoms of the generally undrained depressions scattered irregularly among these hills are at altitudes of only 130 to 160 feet, thus giving a local relief of approximately 125 feet. Here the eastern face of the moraine is a somewhat broken scarp, 30 or 40 feet high, that rises abruptly from the surface of the Mashpee pitted plain at altitudes of 200 to 220 feet. In contrast the western edge of the moraine is represented by a much more sinuous line that generally falls between the 100- and the 120-foot contour line. Six or eight miles southward, near the southwest corner of the Pocasset quadrangle, the higher morainic knobs reach only 150 to 190 feet above the sea, the foot of the east-facing scarp has descended to an altitude of scarcely more than 100 feet, and the western edge of the moraine is generally between 40 and 60 feet above sea level. Nevertheless, the local relief within the morainic belt is just as great as it is farther north; depressions between the hills descend to altitudes of only 40 to 70 feet, and the water surface of

Turtle Pond, in the deepest of the holes in this portion of the moraine, is only 22 feet above sea level. Still farther southward, in the Woods Hole quadrangle, the moraine forms a peninsula jutting southwestward between Buzzards Bay and Vineyard Sound. Here its higher hills have summit altitudes between 100 and 125 feet, and the local relief is only slightly less than elsewhere. Here also the gently sloping surface of the Mashpee pitted plain beneath the eastern front of the moraine descends from an altitude of about 40 feet at Falmouth and disappears beneath the sea a mile northeast of Nobska Point.

Throughout the entire length of the Buzzards Bay moraine, from the vicinity of the canal to Woods Hole, the surface is abundantly covered with huge boulders and subangular blocks of rock ranging up to 18 or 20 feet in greatest diameter. Hardly a 5-acre patch lacks an erratic that is not at least 8 feet across. No stones of similar size are present on the pitted plain to the east where the largest rocks at most places are cobblestones or in some places are boulders 2½ to 3 feet in diameter. Moreover, on the lower ground to the west, between the moraine and the bay shore, large erratics are relatively scarce and are apparently confined to certain, sparsely scattered areas, each 50 acres or so in extent, separated from each other by hundreds of acres devoid of rocks larger than a cobblestone.

Except for these conspicuous blocks and boulders the materials composing the Buzzards Bay moraine are almost completely concealed by a dense growth of grass, shrubbery, and trees. Exposures in road cuts and gravel pits must be amplified by the liberal use of mattock, hoe, and soil auger, if one is to secure a true concept of the composition of this moraine. The casual observer would note the many pits, dug in stratified sand and gravel, and would be tempted to conclude that such materials make up the bulk of the hilly zone. Nearly everywhere, however, there is a veneer of till, 3 feet or more thick, that must be removed before the stratified sand and gravel are exposed, and at many places exposures of till are 10 to 14 feet thick. Nearly all the pits excavated in the morainic belt were intended to secure gravel and sand; consequently, where thick till was encountered, digging was promptly abandoned, and the hillside scars have long since been healed by slope wash, frost action, and the growth of vegetation.

One of the best exposures of till in the Buzzards Bay moraine was noted in a recently worked pit in West Falmouth (northeast corner of Woods Hole quadrangle), on the north side of Blacksmith Shop road half a mile east of the state highway. During the summer of 1939 this showed a fresh cut, 8 to 10 feet high, entirely in till. The middle and lower portions consist of firmly compacted gray sandy till containing many subangular blocks and boulders that range up to 3 feet in longest dimension. There was a suggestion of irregular lamination or pseudo bedding as though ice had pressed heavily upon the deposits. The upper portion is rusty brown as a result of oxidation by ground water and is much less resistant to erosion and to excavating tools than that beneath. At one place the oxidized zone is 4 feet thick, where the large roots of a tree have penetrated far into the ground, but elsewhere the depth of oxidation is approximately 2 feet, beneath the few inches of soil.

Such compact till is one of the materials referred to as "hardpan" by the local residents and well-diggers. It is not, however, typical of the Buzzards Bay moraine. At the majority of exposures, even beneath the zone of oxidized material, the till is loose and easily crumbled. This is due to the small amount of clay in its characteristically sandy matrix or to the lack of compression by overriding ice. In general, the glacialist familiar with the drift of the Mississippi Valley and New England would refer to the nonstratified material ordinarily observed in this moraine as "sandy till" rather than as "boulder clay."

The combination of topographic features, distribution of large erratics, and scattered exposures of fairly thick bodies of till leads to the conclusion that this is a true glacial moraine, not a "pseudo-moraine" as that term is used by some glacialists. It marks a former position of the margin of the ice sheet and therefore may be appropriately referred to as a frontal moraine.

Some of the depressions in this moraine are doubtless due to the tardy melting of buried ice blocks. Others presumably result from the irregularity of the surface on which the ice was resting. The great majority, however, are irregular hollows where the total thickness of glacial debris was less than that composing the surrounding hills. A few of these hollows have been partly filled by glacio-fluvial sand and silt, and presumably there has been at least a little slope wash and soil creep in all of them. Many contain bogs, several sufficiently large and suitably located to be used for the cultivation of cranberries. In some there is a considerable accumulation of peat. A very few of the depressions near the western edge of the moraine contain permanent ponds; subsurface drainage is apparently competent to remove the rain water from any hollow, the bottom of which is not less than 50 feet above sea level.

One of the larger of the shallow, flat-floored depressions, apparently the site of considerable glacio-fluvial or glacio-lacustrine deposition, is located three-eighths of a mile southeast of Signal Hill, near the northern end of the Buzzards Bay moraine, 1½ miles northwest of the center of the Pocasset quadrangle. Here a conspicuous treeless glade at an altitude of 200 feet is traversed by Signal Hill road and surrounded by irregular wooded hills that rise 50 feet or more above the tiny plain. The soil auger revealed that beneath a surface zone of chocolate-brown to black sandy soil, 10 inches thick, there is a zone of rusty sand 2 feet thick, and beneath that 15 feet of clean, yellow, medium to coarse sand with a few layers containing pebbles that range up to 2 inches in diameter.

In contrast with such places where the sand and gravel presumably overlie till, there are many localities where the relations of till overlying silt, sand, and gravel are clearly displayed. The walls of a recently excavated pit on the south side of the Pocasset-Forestdale road, one mile east of the state highway, near the center of the Pocasset quadrangle, were free of slumped material during the summer of 1939. Beneath the soil zone, 4 or 5 inches thick, with whitish, bleached humus at its base, there is 18 to 33 inches of light-tan, oxidized till containing many stones 4 inches or more in diameter and subangular blocks a foot or more in length. This till rests upon laminated silt or fine sand with many layers of sticky clay, containing a few, presumably ice-rafted, cobblestones and pebbles. The uppermost 4 to 10 inches of this water-laid deposit is oxidized to a bright or rusty red; the rest is gray or brownish gray. The upper 2 to 5 feet of this stratified material is contorted into folds, some of which are overturned and broken by thrust faults, as though the whole had been shoved from the west by overriding ice. An auger hole sunk in the floor of the pit, beginning 9 or 10 feet below the top of the waterlaid silt, sand, and clay, revealed the fact that the excavation had stopped at the bottom of this type of material. A few inches below the pit floor the auger encountered gray till, not quite so compact as that described in a preceding paragraph as seen in the lower part of the pit near the Blacksmith Shop road, but otherwise similar to it, and continued in till to a depth of a little more than 3 feet. Beneath the till was another zone, less than a foot thick, of laminated silt and clay, resting upon sandy gravel or sandy till into which the auger could be driven to a depth of only 2 feet.

At this locality there seems to be no valid reason for postulating any long interval of exposure of the laminated silt and clay to the agents of weathering and oxidation prior to the deposition of the overlying till. The observed phenomena may well have resulted from oxidation proceeding from the surface downward through the till to affect the uppermost few inches of the stratified materials after the final withdrawal of the ice. If that interpretation is correct, the whole deposit of intercalated till and glacio-fluvial or glacio-lacustrine materials may well have been laid down close to the edge of a slightly fluctuating body of ice, the margin of which at times was a fraction of a mile to the west and at other times advanced a similar distance to the east. In the last such oscillatory advance, the moving ice distorted the laminated silt, sand, and clay over which it rode. It is precisely the sort of heterogeneous deposit that one would expect to see in a frontal moraine constructed beneath the edge of a glacier, the margin of which shifted from time to time across a belt a mile or so wide.

TABLE 1—*Mechanical analysis of material from Brick Kiln Road pit*

(Mm.)	(Per cent)
0.3 and over	0.2
0.25–0.29	0.3
0.15–0.24	0.1
0.10–0.14	0.7
0.04–0.09	20.0
0.02–0.03	20.0
0.01–0.02	24.0
Less than 0.01	34.0

The till and associated water-laid materials are also exposed in the "claypit" on the north side of Brick Kiln road, three quarters of a mile east of the state highway, just within the west margin of the Falmouth quadrangle. Here there is an irregular area of flat land, about 120 feet above sea level, covering 40 or 50 acres, approximately midway between the east and west boundaries of the Buzzards Bay moraine. At several points the edges of this small plain overlook deep hollows, and at other places it abuts against the slopes of irregularly rounded hills that rise 20 or 30 feet above it. The broad shallow pits in the western part of the flat area were dug to secure "clay" for the construction of tennis courts for which the material seems to be peculiarly suitable. Apparently it was also used, years ago, for the manufacture of brick.

The "clay" is in reality a finely laminated silt with alternating layers of reddish-brown and brownish-gray sediment that in some places recur so regularly as to resemble varves. Some of the grayish laminae have the fine texture and sticky consistency of clay, whereas the reddish laminae are coarser and less plastic. Individual layers are commonly only a few millimeters thick. A mechanical analysis (Table 1) of a sample taken 6 feet below the top of the deposit and containing material of both types of laminae was made by Arthur S. Knox.

Thus it would appear that approximately two-thirds of the particles range between 0.01 and 0.10 of a millimeter in size, whereas about one-third of them are less than 0.01 of a millimeter in diameter.

At both the north and south ends of the excavation the laminated silt is overlain by till, 2 or 3 feet thick and containing the customary assortment of cobblestones and blocks ranging up to 2 feet in length. Several larger erratics are scattered over the floor of the abandoned portion of the pit, and others appear on the slopes rising beyond the flat area. Where overlain by till the upper foot or two of the laminated silt is slightly arched into sinuous folds of varying dimensions; at only one place, as exposed in the summer of 1939, were these folds sheared or mangled (Pl. 1, figs. 1, 2). In general the uppermost 2 to 2½ feet of the silt is coarser, sandier, less finely laminated, and more irregularly bedded than the lower portion. Surface oxidation has proceeded irregularly downward from the base of the soil zone to depths of 2 or 3 feet and is much more conspicuous in the coarser laminae than in the fine. At certain points, near the north end of the pit, the lower 2 to 5 inches of the overlying till, where the surface of the silt is lower than on either side, is thoroughly oxidized to a deep red.

At one point, also near the north end of the pit, as developed during the summer of 1939, a glacial erratic more than a foot long was embedded 2½ feet below the surface of the laminated silt (Pl. 1, fig. 3). The laminae were bent downward beneath the erratic, conforming to its lower surface, but were horizontal a few inches above its upper surface. It was unquestionably a berg-rafted stone dropped to the floor of the pond in which the silts were accumulating at the time of their deposition. Probably some of the erratics left on the floor of the pit during excavation had a similar origin.

The maximum thickness of the exposed silt was 8 feet. An auger hole in the floor of the pit at that spot penetrated 7 feet more of the same sort of material and continued downward through 5 feet of clean, well-washed sand and gravel. The vertical section of the material at that point may therefore be given as follows:

DEPOSITS AT BRICK KILN ROAD, NORTH OF FALMOUTH

Soil	4 inches
Till; sandy, brownish	2 to 3 feet
Silt; laminated, contains berg-rafted boulders	15 feet
Sand and gravel	5 feet +

Another auger hole near the center of the flat area east of the "claypit" revealed the following sequence of materials:

DEPOSITS NEAR BRICK KILN ROAD PIT, NORTH OF FALMOUTH

Soil, with humus and roots	6 to 8 inches
Soil; rusty and sandy	1 foot
Silt and clay; laminated like that exposed in the pit, 300 yards to the west; dark, rusty sand streaks at about 8 and 10 feet. Boring stopped by pebbles in rusty sand at 11.5 feet	10 feet

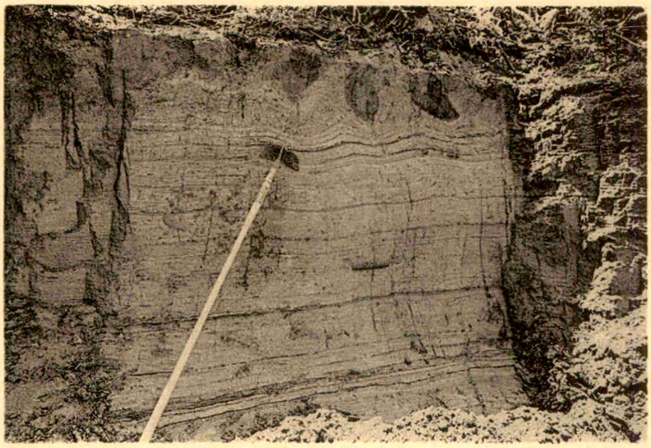


FIGURE 1. LAMINATED SILT EXPOSED IN SUMMER OF 1939 AT SOUTH END OF PIT

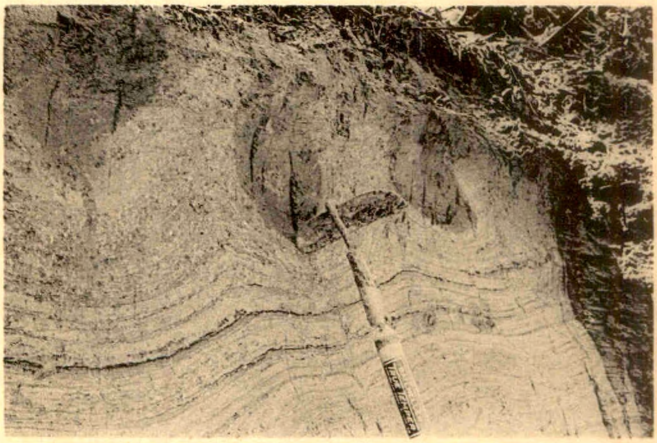


FIGURE 2. UPPER PORTION OF SAME EXPOSURE AS IN FIGURE 1 Showing lumps of till pressed into contorted laminae of silt, sand, and fine gravel.

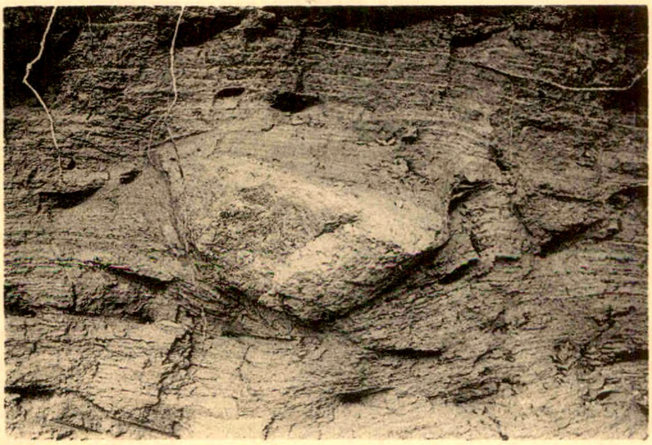


FIGURE 3. BERG-RAFTED BOULDER APPROXIMATELY 20 INCHES LONG Embedded in laminated silt and clay at north end of pit, as developed in Summer of 1939.

GLACIO-LACUSTRINE DEPOSITS IN BRICK KILN ROAD PIT

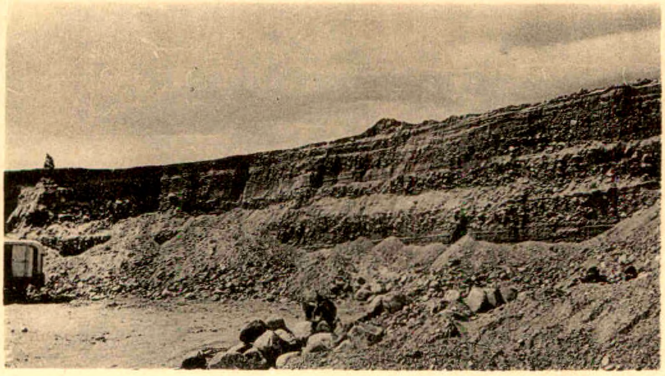


FIGURE 1. COARSE GRAVEL EXPOSED IN LAWRENCE COMPANY PIT
West of Sols Pond in Falmouth.

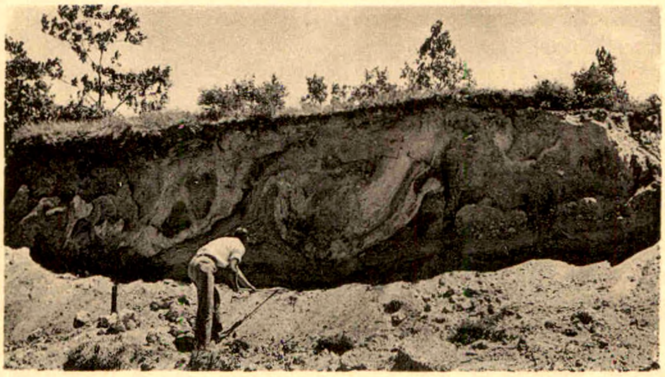


FIGURE 2. LARGE CLUMPS OF TILL EMBEDDED IN STRATIFIED SAND AND GRAVEL
In this exposure at southeast side of gravel pit near Teaticket, both the till clump in middle of vertical face and the one directly above the man are irregularly stained with iron oxides.

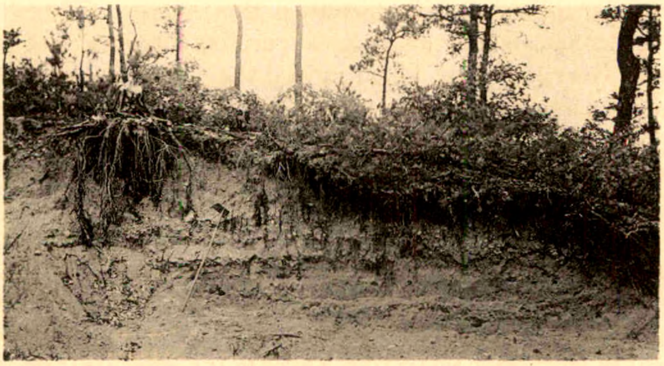


FIGURE 3. HORIZONTALLY BEDDED SAND AND GRAVEL IN ROAD CUT ½ MILE SOUTHEAST OF JENKINS POND ON ASHAMET-GRASSY POND ROAD
Bedding is truncated by gently sloping wall of a furrow trending nearly at right angles to road cut.

MATERIALS COMPOSING MASHPEE PITTED PLAIN

Samples of the laminated silt were collected in the pit at depths of 4, 6, and 8 feet below the overlying till by Arthur S. Knox for a study of the microflora which they might contain. He reports that each of the three samples rated 2,21d on the Ridgway color scale and contained unrecognizable carbonized fragments that might be remains of seaweed and unidentifiable yellow resinous bodies. No pollen was observed, but all three contained the fossils listed below, although not all the varieties of sponge spicules were noted in each:

Fossils in Silt From Brick Kiln Road Pit

(identified by Arthur S Knox)

Diatoms—very rare (marine or brackish water)
  Melosira sulcata (Ehrenberg) Kutzing
  Coscinodiscus radiatus Ehrenberg
Sponge spicules—mostly broken (marine or fresh water)
  Oxea megasclera, 204 to 306 by 11 9 to 17 microns
  Styli monactinal megasclera, entirely spined, 230 by 11 9 microns
  Styli monactinal megasclera, smooth, up to 510 by 17 microns
  Fusiform acerate megasclera
  Clavate megasclera
  Large spicules, probably triaene tetraxonids (plagiotriaene), up to 102 microns in diameter
  Sterraster microsclera (fragments)

The occurrence of marine or brackish-water diatoms and marine or fresh-water sponges, all of which are Pleistocene or Recent forms, in these sediments at an altitude of more than 100 feet, and beneath the glacial till that veneers the surface of the Buzzards Bay moraine, makes it necessary to reconsider the views of Shaler concerning this major feature of Cape Cod geology. He believed (1898, p. 554-559) the ridge-like appearance of that portion of the "Cape Cod moraine" here designated as the Buzzards Bay moraine was not a result of glacial deposition but that the glacial debris was merely a thin veneer resting "on the summit of a pre-Glacial divide which separated the waters of the old Buzzards Bay river from that which formed the valley that is now Vineyard Sound." Although there is no evidence that Shaler had seen the laminated silts at or near the locality just described in detail, those sediments would undoubtedly fall within the category of "pre-Glacial rocks" described by him as composing "the Falmouth continuation of the Plymouth ridge."

The argument as Shaler might present it in the light of present knowledge would be somewhat as follows: During an early Wisconsin substage of glaciation, or better during a pre-Wisconsin Pleistocene glacial stage, an ice margin stood somewhere to the north of the Falmouth quadrangle, the whole of which was at that time much lower in relation to sea level than it is today. Sand, silt, and clay accumulated on the sea floor, possibly in an estuarine re-entrant of the sea, and berg-rafted boulders were dropped into the laminated sediments. Diatoms and sponges, inhabiting the marine or brackish water at this place, contributed the fossils that have been discovered. Later uplift raised the sea floor at least 120 feet above sea level at this spot. Subsequently erosion removed much of the marine strata and left a high hill or ridge on the divide between the two "pre-Glacial" valleys that he mentions. Later still, the ice of some one of the Wisconsin substages advanced on the hill without greatly changing its topography and on retreat left it veneered with a thin irregular coating of till. Two alternatives present themselves, the first of which probably, though not certainly, would be approved by Shaler. (1) The marine silts might be regarded as much older than the moraine along the northwest shore of Martha's Vineyard, and therefore the hill was overridden by ice that reached that far beyond it toward the south; (2) the marine silts might be interpreted as formed after the ice margin had withdrawn far to the north from Martha's Vineyard and that therefore the interval of time between the deposition of the moraine on Martha's Vineyard and the glacial readvance to the Buzzards Bay moraine was long enough to permit a major shift in the strand line and a very considerable amount of subaerial erosion.

Neither of these hypotheses seems tenable. If the first is accepted, the deep hollows in the Buzzards Bay moraine in close proximity to the laminated silts would imply that a flat-topped hill rising steeply 50 feet or more above its surroundings and composed of unconsolidated silt and sand had withstood erosion by ice which overrode it and advanced 15 or 20 miles beyond it. If the second hypothesis is adopted, the close similarity in amount of weathering, oxidation by ground water, and modification by stream erosion that characterize the two moraines in question would seem to indicate that the time between their deposition was altogether inadequate for the large amount of subaerial erosion required.

A third hypothesis might be constructed on the theory that the ice advancing to the site of the Buzzards Bay moraine shoved the laminated silts forward and upward from their original position to their present location. This would make unnecessary any appeal to an unreasonably long interval of erosion between the deposition of the two moraines. It is, however, opposed by the fact that the silty beds appear to be quite horizontal throughout a considerable area, much too extensive to be interpreted as the flat summit of a broad anticlinal fold or as part of an overthrust block composed of such incoherent material.

Such considerations lead to the belief that the laminated silt and underlying sand were originally deposited during the construction of the moraine of which they appear to be a part. Melt water from the ice responsible for the Buzzards Bay moraine was locally ponded in a hollow close to the irregularly frayed margin of the ice, perhaps with tongues or blocks of stagnant ice on either side. Here the sand and silt transported by the melt water filled the depression and built a small but ordinary "sand plain." Icebergs drifting across the pond dropped berg-rafted boulders in the midst of the accumulating sediment. Fossil remains of diatoms and sponges that had lived in some previous interglacial interval in the waters then occupying the present site of Massachusetts Bay to the north were transported in the ice, with the marine clay in which they were embedded, from some place where the advancing glacier plowed into the sediments on the shallow sea floor. Liberated when the ice melted, these fossils were washed into the pond by the melt water along with the particles of silt and clay, just like any other glacial erratics. Later a slight readvance of the ice shoved its thin margin over the west edge of the sand plain, disturbing the uppermost layers of silt and at last depositing the thin veneer of till that overlies the silt in the vicinity of the "claypit." Final disappearance of the stagnant blocks and tongues of ice, beneath or beyond the active ice that participated in that final movement, left the deep holes near the silt plain and the ice-contact-slope along its northeastern margin.

That the ice responsible for the Buzzards Bay moraine moved southward along a route involving the traverse of a portion of the area now occupied by the waters of Massachusetts Bay is indicated by the variety of stones in the till of that moraine. During what stages of the Pleistocene the portion of the continent immediately east of Boston was submerged and for how long the submergences continued are unknown; but the presence of occasional diatoms and fragmentary sponge spicules in the till at several places in Cape Cod, east of the Buzzards Bay moraine, indicates that the ice had access to marine clay containing such fossils and that some of them survived the vicissitudes of transport in the ice.

In the absence of bedrock suitable for the recording of glacial movement, the only basis for discovering the direction in which the ice advanced is that provided by the identification of the transported stones in terms of their known sources in eastern and northern New England. "Stone counts" were therefore made at a dozen localities in various parts of the Buzzards Bay moraine. At each exposure generally about 200 (at some places only 100) pebbles, between 1 and 3 inches in diameter, were picked up at random, an effort being made to collect all such stones that were available within a radius of a foot or two from some selected point in the midst of a body of till. All the stones in each collection were later classified on the basis of their lithologic

TABLE 2.—*Stone-count data*

Composition (in percent) of pebbles in Cape Cod moraines

Locality*	Buzzards Bay moraine								Sandwich moraine							Area of overlap				
	G-80	G-79	G-33	G-38	G-78	G-14	G-82	Average	G-55	G-42	T-80	G-98	T-82	T-83	Average	T-67	T-124	T-121	T-122	Average
Granite (including gneissoid granite, aplite, pegmatite)	69	72	70	67	61	66	58	66	55	36	38	35	57	50	45	56	56	71	58	60
Quartz-feldspar gneiss	0	6	0	1	0	0	0	1	0	5	2	4	5	3	3	0	0	0	0	0
Diorite and gabbro (including biotite gneiss)	3	1	0	0	9	9	2	3	4	4	7	3	1	2	3.5	1	4	0	3	2
Basalt (including basalt porphyry)	3	2	3	2	3	4	3	3	4	11	11	13	5	13	9.5	6	6	5	3	5
Other volcanics (mostly rhyolite and pyroclastics)	2	2	0	6	6	5	8	4	10	18	16	19	9	10	13.5	15	11	5	12	11
Schist (including greenstone)	7	7	0	3	1	1	2	3	0	3	4	5	9	9	5	3	3	7	5	4.5
Slate (including argillite)	0	0	0	1	0	0	0	t	2	3	7	4	3	6	4	0	4	2	2	2
Quartzite	12	6	18	12	15	5	11	11	14	5	10	10	4	2	7.5	9	8	6	14	9
Vein quartz	5	4	9	8	6	7	16	8	11	11	4	6	6	6	7	7	9	6	4	6.5
Arkose and sandstone	0	0	0	0	0	0	0	0	0	3	1	2	0	0	1	1	0	0	0	t
Conglomerate	1	0	0	0	0	0	1	t	0	0	0	0	0	0	0	0	0	0	0	0
Miscellaneous	0	1	0	1	0	3	1	1	1	3	2	0	1	0	1	2	0	0	1	1
	Southwest			to			Northeast		West			to			East					

* Localities designated G represent collections by Goldthwait; T, Thiesmeyer.

character. The results obtained by the more significant of these "stone counts" are listed in Table 2. At the same time, specific identification of stones derived from many of the known formations of eastern Massachusetts, southern New Hampshire, and southern Maine was attempted. Valuable assistance in making such identifications was rendered by L. W. Currier, Marland P. Billings, Harry Berman, and other geologists familiar with the geology of those areas.

By far the most abundant rocks among the erratics of this moraine are the granites and their close allies. Most of these were undoubtedly derived from the Dedham granodiorite and associated rocks of the region adjacent to the Boston basin. Others closely resemble certain of the granites exposed in southern Maine. Diorite, of the type known to occur near Salem, Massachusetts, and diabase, identified as having been derived from the widely scattered system of dikes named for Medford, are fairly common. Volcanic rocks of the types occurring at Nantasket and Lynn comprise a small but significant fraction of the assemblage. As noted by Howe (1936, p. 394-396) the fragments of the distinctive felsite known to occur in place only in the northern part of the town of Hingham, Massachusetts, are sparingly scattered throughout the Buzzards Bay moraine. On the other hand, no specimen of granite that could be identified as having been derived from Quincy granite was found. Among the few fragments of sedimentary rock are Roxbury conglomerate and Cambridge slate, the most widespread of the formations in the Boston basin. A few small fragments of laminated, nonfissile slate, like the varved slate of Squantum, Massachusetts, were also noted. Thus, it would appear that ice moving southward from southern Maine and southeastern New Hampshire across the eastern margin of Massachusetts could have gathered all the diverse materials found in the Buzzards Bay moraine.

SANDWICH MORAINE

A belt of hummocky and relatively high terrain, known to the local residents as "the backbone of the cape," stretches eastward from the Cape Cod canal across the northern part of the Pocasset and Sandwich quadrangles. Woodworth considered this a part of the Falmouth moraine, contemporaneous in origin with the other portions here designated as the Buzzards Bay moraine. It is, however, a distinct moraine, differing significantly from the southwestward-trending moraine in composition, in relations to the lobate margin of the ice sheet, and in time of emplacement. The name Sandwich moraine is therefore applied to it, inasmuch as it traverses for many miles the northern part of the town bearing that name.

Driving east from Sagamore to Dennis on the principal highway, one has constantly on his right the irregular skyline of this "backbone" of Cape Cod. In the Sandwich quadrangle, the higher summits stand between 200 and 300 feet above sea level; of these, Telegraph Hill, with its summit altitude of 292 feet, one mile south of Sandwich village, is the highest point. Eastward along the moraine the hilltops decrease somewhat in altitude so that in the belt just south of West Barnstable the higher summits are between 150 and 200 feet above sea level. Across most of the Sandwich quadrangle the moraine is more than a mile in width but, at the eastern margin of that quadrangle, it narrows to half a mile.

The surface of this morainic belt is everywhere characterized by irregular hills and undrained depressions that give a topographic expression quite like that of the Buzzards Bay moraine. In the area extending a mile or so east and southeast from Telegraph Hill, many of the small hills are elongated parallel to the trend of the moraine. Presumably these rude ridges are the result of deposition of frontal debris or of the pushing forward of marginal drift during minor fluctuations of the ice front.

The south face of the Sandwich moraine overlooks the Mashpee pitted plain along a well-defined boundary. South of Telegraph Hill and along Popple Bottom road the moraine front appears like a high ridge strewn with scattered erratics. The moraine boundary is, however, complicated just north of the Cape Cod Airport in the southeast part of the Sandwich quadrangle by subsidiary short ridges of bedded gravel which extend obliquely into the main plain (Fig. 3). As explained in the discussion of the Mashpee pitted plain. these were probably formed near the ice edge prior to the deposition of the main moraine.

The northern part of the Sandwich moraine is very irregular, commonly sandy, and low, like the west side of the Buzzards Bay moraine. There is, however, a marked topographic break along the north boundary of the Sandwich moraine, even more conspicuous and continuous than that along the west side of the Buzzards Bay moraine. North of the Sandwich moraine lie areas covered with salt marsh, low sand plains, and a few distinct and entirely separate hills of the Scorton moraine (Fig. 2), described on a subsequent page.

At most of the exposures the material in the Sandwich moraine is sandy till. A representative pit, on State Highway Number 130, opposite the entrance to the Telegraph Hill fire tower, shows a 10-foot thickness of till, including a few big angular blocks and a few clumps of well-sorted sand. These sand clumps may be the remains of frozen masses picked up by the invading ice. Hard-packed, gray till or "hardpan," believed to be of the same age as the loose, sandy till elsewhere in the moraine, is exposed behind the farm at the north end of Gully Lane east of Sandwich Village and at a few other places.

Although gravel pits and deep road cuts are not so numerous in the Sandwich moraine as in the Buzzards Bay moraine, there are enough such exposures to indicate that the former, like the latter, is essentially a veneer of till, 5 to 30 feet or more thick, resting on disturbed glacio-fluvial sand and gravel. Road cuts, for example near the north edge of the moraine along Quaker Meetinghouse road and Chase road, reveal sandy till underlain by bedded sand.

Large erratics are generally less numerous on this moraine than on the Buzzards Bay moraine. In the region north of Telegraph Hill or along Maple Swamp road near the center of the Sandwich quadrangle a person may travel almost a mile without seeing one. On the other hand, many areas in the southern part of the moraine are liberally sprinkled with large blocks. Along Forest Street, south of West Barnstable, and north of Popple Bottom road, northeast of Farmersville, many boulders and blocks are 2 to 20 feet long. The largest single erratic in this moraine, exclusive of the area of overlap on the Buzzards Bay moraine, may be seen one mile south of East Sandwich; it measures approximately 30 by 15 by 10 feet.

Explanation of the relative scarcity of boulders may be found in the types of rocks represented. Whereas more than 95 per cent of the large erratics in the Buzzards Bay moraine are granite or granite gneiss, from 5 to 20 per cent of the larger stones in the Sandwich moraine are composed of basalt or other volcanic rock. Granite ledges are notorious for yielding large blocks wherever they are crossed by moving ice. We may infer that as it approached Cape Cod the lobe of ice that formed the Buzzards Bay moraine passed over a larger area underlain by granite than did the ice that formed the Sandwich moraine.

The composition of the pebbles, 1 to 3 inches in diameter, also indicates a significant difference in the till of the two moraines. Fragments of bright-red, arkosic sandstone and conglomerate, finely laminated black and white quartzite, and black slate are common in the Sandwich moraine, whereas none of these rock types is found in the Buzzards Bay moraine Moreover, the identification of the rock types represented among 200 pebbles of such sizes collected at random at each of six localities in the Sandwich moraine shows different percentages of the types that are common to both moraines. (*See* Table 2.) On the average only 45 per cent of the small stones in the Sandwich moraine are granite, and the maximum for these collections is 57 per cent, whereas in the Buzzards Bay moraine the average is 66 per cent, and the minimum is 58 per cent.

Basalt pebbles make up nearly 10 per cent of the smaller stones in the Sandwich moraine but only 3 per cent of those in the Buzzards Bay moraine. Other volcanic pebbles average 13 per cent in the Sandwich moraine and only 4 per cent in the Buzzards Bay moraine. Of course, the pebble content of any moraine changes from mile to mile of its length, but here the difference in rock content can be traced to within a few hundred feet of the boundary between the two moraines.

Doubtless, the sources from which the ice secured the materials that were deposited during the construction of the Sandwich moraine are now for the most part concealed beneath the waters of Cape Cod Bay and the Gulf of Maine. The fragments of red sandstone and conglomerate, many of which were noted in the gravel pit just west of highway 130, are particularly interesting in this connection. Lithologically, these specimens resemble the Triassic redbeds in every respect. Their most likely source is in the Triassic basin extending southwestward from the head of the Bay of Fundy and largely submerged beneath the Gulf of Maine. As interpreted by Johnson (1925, p. 295) the Triassic rocks extend along the sea floor to a point only 100 miles northwest of the Sandwich moraine.

RELATIONS BETWEEN SANDWICH AND BUZZARDS BAY MORAINES AND MASHPEE PITTED PLAIN

The Mashpee pitted plain and its origin are discussed in detail on subsequent pages, but it is desirable to summarize briefly here the conclusions that have been reached concerning the relations between the two large moraines of western Cape Cod and the adjoining outwash plain. The data now available indicate that the Mashpee pitted plain was essentially constructed by meltwater issuing from the re-entrant angle between the Buzzards Bay lobe and the Cape Cod Bay lobe of the Laurentian ice sheet at a time when its margin stood some distance northwest of the Buzzards Bay moraine front and north of the Sandwich moraine front. After the greater part of the sand and gravel of the plain had been deposited, both lobes of ice pushed forward, plowing into and overriding the previously deposited outwash. Thus the frontal moraines were deposited on the edges of the fan-shaped plain. Meltwater issuing from the ice while these two moraines were under construction contributed relatively little debris to the outwash plain.

At several places, large blocks of stagnant ice buried in the sand and gravel of the plain were overridden by the readvancing ice and remained intact until after the moraines had been constructed and the lobate margin of the ice sheet had again retreated northwestward and northward. When these buried ice masses finally melted, large kettle holes were formed, partly in the moraine and partly in the adjacent plain

(Fig. 3). Thus the large depression traversed by Forestdale road, 0.7 mile southwest of Telegraph Hill in Sandwich, extends from the Sandwich moraine into the pitted plain. A similar origin is indicated for the long, irregular kettle hole, 1.4 miles due east of Halfway Pond in the northern part of the Pocasset quadrangle, the western half of which is within the Buzzards Bay moraine, the eastern half within the Mashpee pitted plain. Likewise, Long Pond, 1.5 miles north of Falmouth, and Oyster Pond, 1.8 miles northeast of Woods Hole, occupy kettle holes partly within the Buzzards Bay moraine and partly in the Mashpee pitted plain.

OVERLAP OF SANDWICH MORAINE UPON BUZZARDS BAY MORAINE

The Sandwich and Buzzards Bay moraines coalesce in the area south of the Cape Cod canal and extending from the Sagamore into the Pocasset quadrangle. (*See* Figure 2.) The data observed here indicate that the Buzzards Bay lobe of ice had begun to retreat before the Cape Cod Bay lobe had completed its readvance to the position marked by the Sandwich moraine. Consequently the west end of that moraine overlaps the north end of the Buzzards Bay moraine.

This area of overlap is as irregular and knobby as any parts of either moraine that lie to the south or east. The morainic topography extends northwestward across the canal as far as Great Herring Pond and includes the Bournedale Hills, traversed by U. S. Highway No. 6 as it approaches the Sagamore Bridge from the west. The highest point here is Signal Hill, 211 feet in altitude. Many large granite boulders are strewn over this portion of the Sandwich moraine and lie in clusters along the transmission line on either side of the canal and in the Shawme State Forest. The largest of these, Sacrifice Rock, half a mile northwest of the fire tower on Signal Hill, has split into several parts that lie against one another, but portions of its surface retain the marks of sandblasting. Its dimensions originally were at least 30 by 20 by 15 feet.

Cuts along the transmission lines and adjacent country lanes have exposed sticky, tan to brown till with abundant buried erratics of granite and other rocks. Several pits at different places disclose stratified gravel associated with till. On Canal View road, both north and south of the boundary between the Pocasset and the Sagamore quadrangles, there are pits in coarse, stratified gravel that include water-worn boulders, a few of which are as much as 4 feet long. When the Cape Cod Canal was excavated through a low portion of this area of overlap and along the natural drainage line leading southwestward from Great Herring Pond to Buzzards Bay, it exposed sandy till with huge embedded erratics overlying stratified sand and gravel. Here as elsewhere in the Sandwich and Buzzards Bay moraines, till was spread over earlier glacio-fluvial deposits, and some glacio-fluvial deposition accompanied its formation.

Certain types of pebbles found in the Sandwich moraine but not observed in the Buzzards Bay moraine are present in this area of overlap. These are chiefly chips of black slate, pink quartz-feldspar gneiss, red arkose, and laminated quartzite. The four pebble counts (*see* Table 2) made in this area indicate that the proportions of basalt, quartzite, and vein quartz are intermediate between those found to be characteristic of the rest of the Sandwich moraine and of the Buzzards Bay moraine. The percentage of granite pebbles resembles that in the Buzzards Bay moraine, whereas the percentage of volcanic pebbles is like that of the Sandwich moraine. This may be explained adequately as the result of overriding of the northern end of the Buzzards Bay moraine by southwestward-moving ice of the Cape Cod Bay lobe. As the ice front fluctuated back and forth in a fairly narrow belt, its debris was mixed with that into which it plowed or over which it advanced.

The southern limit of the till displaying these lithologic characteristics indicative of mixed origin is a rather definite line just south of Kendrick road near the north margin of the Pocasset quadrangle. This line also marks a notable topographic change, with higher hummocks and ridges lying to the south. Thus defined, the belt of what appears to be the westward extension of the Sandwich moraine cuts across the trend of the Buzzards Bay moraine and northwestward for a short time.

Banked against this westernmost edge of the Sandwich moraine, in the area immediately east of the traffic circle at the south end of the Bourne Bridge, there are unusually thick bodies of stratified sand and gravel, including many very large boulders. These constitute the apex of a narrow fan of outwash, spread by meltwater from the Cape Cod Bay lobe that escaped southward between the Buzzards Bay moraine and the front of the Buzzards Bay ice lobe when that ice had withdrawn 2 or 3 miles back from the moraine that it had just constructed. The nature of these deposits west of the Buzzards Bay moraine has been adequately discussed elsewhere (Mather, *et al.*, 1940, p. 40-43), and reference to them is made here merely to cite the additional evidence they provide for the conclusion that the deposition of the Sandwich moraine was completed somewhat later than that of the Buzzards Bay moraine.

SCORTON MORAINE

Between the Sandwich moraine and the shore of Cape Cod Bay there is a discontinuous line of elongate hills to which the name Scorton moraine may appropriately be applied. (*See* Figure 2.) Spaced at unequal intervals from Sagamore Village through Sandwich to Barnstable, these hills include the hill at the northern edge of Sagamore Village, Sagamore Hill, Town Neck Hill, Spring Hill, Scorton Neck, and Scorton Hill. As a rule, each hill displays a broad, flat summit characterized by irregular hummocks and small kettle holes as much as 25 feet deep. Large erratics are sprinkled over each segment of the moraine. As in the Sandwich moraine, these are neither closely packed nor very abundant, but some are of remarkable size.

The structure of these hills is characteristically displayed in a pit near the highway at the south side and near the west end of Scorton Neck. At this locality, sandy till, 6 to 8 feet thick and including irregular masses of sand, rests on stratified sand containing thin seams of gravel. Again, on the hummocky top of Spring Hill, a mile northwest of East Sandwich, a small pit exposes compact till. The soil auger disclosed that this till rests on stratified sand a few feet below the floor of the pit. There are similar indications that Scorton Hill and the hill north of Sagamore Village are also underlain by sandy gravel and veneered with till.

The manner in which the ice deposited this discontinuous moraine is not yet clear. Further studies in adjacent areas must precede final inferences concerning the origin of the moraine and its associated deposits, but some conclusions and tentative working hypotheses can be developed from the information now available.

Most likely the Scorton moraine was never a continuous belt like the Sandwich moraine. If the ice retreated from the Sandwich moraine largely through downwasting and consequent thinning, a very ragged margin would be produced along which large masses of ice may have been quite stagnant. Between these and in the re-entrant angles along the ice front, glacio-fluvial gravel and sand would be deposited by the meltwater. Readvance of thicker portions of the ice sheet from the north would push forward the deeply indented, sawtooth edge of the Cape Cod Bay lobe of ice to override the masses of inactive, debris-clogged ice along its margin and the bodies of glacio-fluvial debris that had been deposited between them. Till would then be piled on top of the patches of stratified sand and gravel. Later, when all the ice had melted away, hills of gravel capped with till would remain where the thicker bodies of gravel had been deposited, and the lower ground near-by where stagnant ice had stood would be covered unevenly with thin till. Still later the meltwater from the dwindling ice at the north might spread sand and sandy gravel over these lower areas and blanket the till. The area would thus consist of isolated, till-veneered hills of gravel standing like islands in a sea of sand This is essentially the condition of the Scorton moraine today.

ELLISVILLE MORAINE

A belt of very irregular topography, the surface of which stands generally higher than the adjacent countryside and is dotted with thousands of large erratics, extends northwestward across the northern portion of the Sagamore quadrangle from the shore of Cape Cod Bay near Ellisville. (*See* Figure 2.) The numerous exposures of sandy till, the many knobs and kettles, the large erratics, the linear plan of this hilly terrain, and its geographic relations to other features of the landscape indicate that it marks a temporary halt in the recession of the ice responsible for the moraines that it had earlier deposited south and east of the Cape Cod Canal. It may therefore be appropriately designated as the Ellisville moraine

In contrast to the south and east fronts of the Sandwich and Buzzards Bay moraines, there are few places with distinct breaks in topography along the southwest margin of the Ellisville moraine. Instead, the boulder-strewn moraine merges along an irregular line with the much-pitted northern edge of a large outwash plain that extends southward and southwestward to the head of Buzzards Bay at Wareham. This may be designated as the Wareham pitted plain and displays many features quite like those of the Mashpee pitted plain. Its surface, however, rises to the higher levels of the Ellisville moraine, and in places, as at flat-topped Mountain Hill, a mile southwest of Ellisville, it stands much higher than the immediately adjacent moraine to the north or northeast. Evidently, the Ellisville moraine and the bordering Wareham pitted plain were formed simultaneously. The moraine was not superimposed on the margins of a pitted plain by readvance of ice, as were the Buzzards Bay and Sandwich moraines, but was constructed when the continuous northward retreat of the ice front was halted for a considerable interval, during which the edge of the Cape Cod Bay lobe fluctuated only within narrow limits within the area now covered by the morainic deposits. Further details concerning the Ellisville moraine and the Wareham pitted plain have been published elsewhere (Mather, Goldthwait, and Thiesmeyer, 1940, p. 46-52).

SUMMARY

Three distinct types of recessional moraines are present in the general vicinity of Buzzards Bay near the west end of Cape Cod. The Buzzards Bay and Sandwich moraines were constructed along the margins of the Buzzards Bay lobe and the Cape Cod Bay lobe of the Laurentian ice sheet during the Wisconsin glacial stage by ice that readvanced from a position north and northwest of these moraines to which it had earlier retreated. Thus the morainic debris was superimposed upon the margins of the Mashpee pitted plain and represents in part the result of ice push as well as the deposit made by melting ice.

The Scorton moraine is a discontinuous series of till-veneered hills marking the ragged edge of the Cape Cod Bay lobe at a time of ice wastage that caused the irregular withdrawal of the ice front from the smooth line of the Sandwich moraine. Slight oscillations of the active ice within the Cape Cod Bay lobe resulted in the overriding of the stagnant ice along its frayed margin and left the morainic hills standing in the midst of glacio-fluvial deposits.

The Ellisville moraine is the least complex of recessional deposits. Continuous northward retreat of the margin of the Cape Cod Bay lobe of ice was temporarily halted. The recessional moraine and the bordering apron of glacio-fluvial debris, the Wareham pitted plain, were contemporaneously deposited.

PART II. MASHPEE PITTED PLAIN

SURFACE FEATURES AND COMPOSITION

The Mashpee pitted plain lies within the angle between the Sandwich and Buzzards Bay moraines and occupies about 100 square miles of land surface on Cape Cod (Fig. 3). Several square miles of sea floor beneath Nantucket and Vineyard Sounds represent its submerged southern margin. The plain is in essence a slightly dissected, subaerially constructed, alluvial fan with its apex at an elevation of 220 feet, close to the place of overlap of one moraine upon the other, and its surface sloping gently to the curving shore line, 10 to 14 miles distant toward the south and southeast. At the east, beyond the Pocasset and Falmouth quadrangles, it coalesces with a similar fan which according to Chute (1939, p. 12) may have developed simultaneously.

The surface of the plain is modified by two types of depressions: (1) irregular pits or kettle holes of all sizes, some of which are as much as a mile in diameter and 100 feet in depth, distributed at random, and (2) long furrows increasing in breadth from a few feet to 1000 feet and in depth from a few inches to more than 30 feet as they radiate southward and southeastward across the lower half of the plain. Woodworth (1934, p. 271) called these linear depressions "creases," and others have referred to them as "valleys," "arroyos," or "channels." The designation "furrow" is, however, adopted here because it is a noncommittal term without any implication concerning their mode of origin.

The surface slope of the entire plain, restored by filling in, in imagination, the pits and furrows, is very gently concave (Fig. 4). From the apex south and southeast for 6 miles, the average gradient is 15 to 20 feet per mile. Indeed, close to the apex some nonpitted portions of the plain incline as much as 25 feet in one mile. The outer and lower half of the plain, on the other hand, has a surface slope of only 12 to 15 feet per mile.

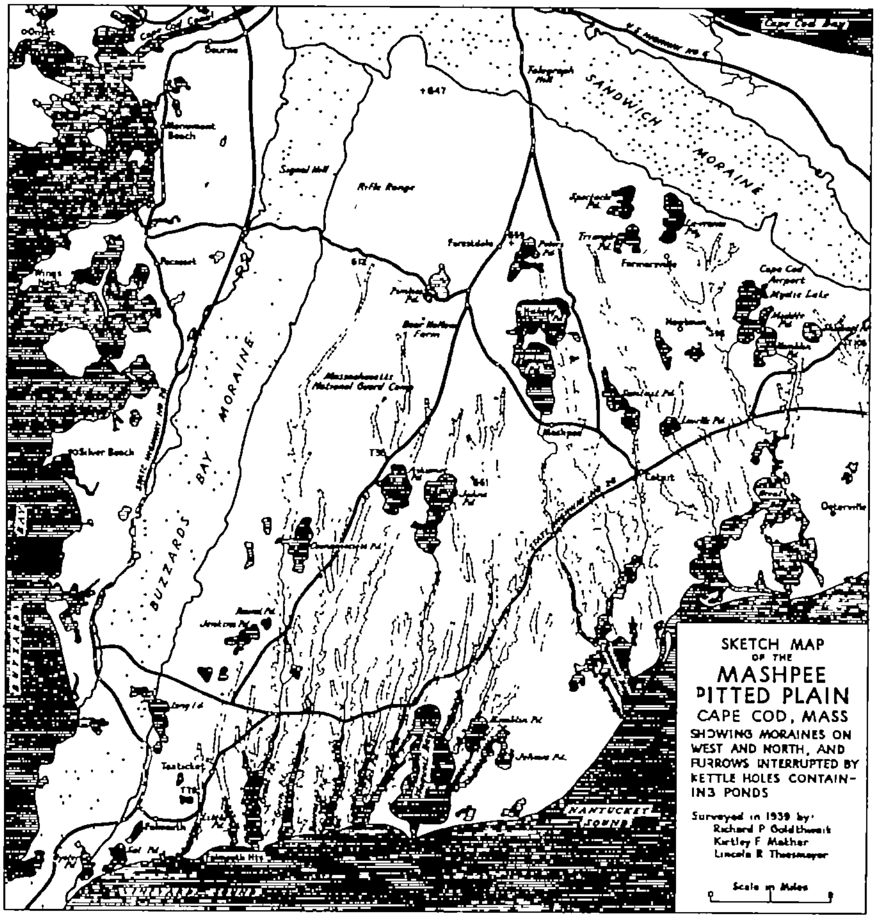


FIGURE 3—*Sketch map of western Cape Cod*

Showing Mashpee pitted plain, bounded on north and west by Sandwich and Buzzards Bay moraines. Furrows traversing plain are delineated by broken lines. Localities at which stone counts recorded in Table 8 were made are indicated by letter and number, *e. g.*, G59.

The material underlying the plain is bedded sand and gravel, accumulated to a thickness of at least 120 feet, the depth of the deepest pits. The average components vary from well-rounded gravel near the apex to subangular sand near the distal edge. Associated with the gravel in the higher portion of the plain are abundant, small, rounded or subangular boulders, 6 to 24 inches in diameter. One pit along the Pocasset road, 1000 feet north of a road intersection, the altitude of which is given on the topographic map of the Pocasset quadrangle as 182 feet, disclosed 4 to 6 feet of very coarse gravel and small boulders overlying 3 feet of cross-bedded sand with thin lenses of gravel. A trench in the Massachusetts National Guard Military Reservation (Camp Edwards) north of Snake Pond and many shallow road cuts in the northern part of the reservation exhibit similarly coarse gravel. In contrast, on the outer, seaward edge of the plain, the many low wave-cut cliffs almost exclusively expose only sand. Pits excavated for sand, used in surfacing roads and bogs, are scattered along the main highway leading northeastward from Falmouth and around each cranberry bog in this vicinity. Very little coarse gravel is available in these pits.

There are exceptions, however, to this gradation in coarseness. Most notable is the Lawrence Company pit, just west of Sols Pond in Falmouth. Although this locality is only 2 miles from the outer edge of the plain, it shows a 6-foot layer of very coarse gravel containing rounded boulders, some of which are as much as 18 inches in diameter (Pl. 2, fig. 1). On the floor of the pit are a few boulders as much as 30 inches long. Similar coarse gravel with rounded cobbles 8 inches in diameter occurs in a near-by pit just southwest of Teaticket. The eastward dip of the layers and slope of the surface between pits in this vicinity suggests that this abnormally coarse material was deposited by meltwater issuing from the adjacent Buzzards Bay lobe while the Buzzards Bay moraine was under construction. Apparently streams from that source built a small local fan on top of the more extensive and older deposits of the Mashpee pitted plain.

The bedding of the sand and gravel composing the fan is essentially parallel to the surface of the plain. Although cross-bedding and channelling are common, no foreset delta structure was observed. The "scour and fill" nature of the deposits is evident at many places where lenticular gravel layers lie between cross-bedded and laminated sands.

The features described above—fan shape, definite concave slope, general decrease of grain size from apex to distal edge, and beds parallel to the surface but vertically variable in texture—are characteristic of deposits of present-day anastomosing streams. The lack of foreset delta structure, even in pits at the distal edge of the exposed plain, indicates that this is a subaerial fan and was not built to accord with any lake surface or sea level.

GLACIAL ORIGIN

The very large natural pits in the plain are most convincing evidence that this is glacial outwash. They occur chiefly in groups near Falmouth, Coonamessett, the rifle ranges in the Massachusetts National Guard

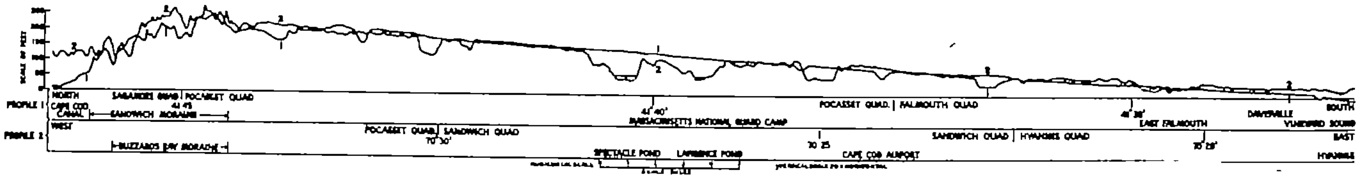


FIGURE 4.—*Profiles across Mashpee pitted plain*

Military Reservation (Camp Edwards), Farmersville, Newtown, the vicinity of Johns Pond, Waquoit, and Osterville. There is no system or regularity to the grouping or to the positions of groups of pits in the plain. More than 30 pits exceed half a mile in diameter. They range in depth from 50 to 120 feet, and several are enclosed completely by gravel walls. The most prominent examples are the depression which contains Snake Pond, west of Forestdale, and the hollows near the center of the Pocasset quadrangle. Some 500 smaller pits are shown as depression contours on the new topographic maps of the United States Geological Survey, and many others observed in the area under discussion are too small to show on maps of the scale used. The side slopes of the pits are inclined 5°, 10° and even as much as 20°. Where closely grouped, the slopes around adjacent depressions may coalesce to form saddle-shaped ridges. The pits vary greatly in shape; some are nearly circular, others elongate oval, and some quite irregular. Within an area of compound pits, the surface may resemble the hummocks and hollows of small "kame and kettle" deposits so common in valleys throughout New England.

These features—the irregular grouping, the great depth, steep slopes, and irregular shape—are prime characteristics of glacial kettle holes. Furthermore, several of these pits, such as those occupied by Long Pond and Oyster Pond near Falmouth, or the elongate depression southwest of Telegraph Hill in Sandwich, extend from the plain into the bordering moraines. As these hollows interrupt the continuous front of the moraine, they must have formed after the moraine was built on top of the outwash plain. Clearly, all these depressions must have resulted from the melting of ice that had been buried during the deposition of the plain.

Two other conditions point to the fact that this is glacial outwash: (1) the plain heads at an altitude of more than 200 feet; river deposits at this altitude could have formed here only during the existence of glacial ice to support the loaded channel of a stream at this high level; (2) fragments of rock of the types known to be in place many miles to the north and west are abundant among the pebbles in the gravel of the plain. Pebbles of the Squantum tillite member of the Roxbury conglomerate, for example, were found at Falmouth Heights and in a pit west of Coonamessett River, northeast of Round Pond. Carboniferous conglomerate pebbles from the Boston or Narragansett Basins were found at a dozen or more exposures. Gray-green melaphyr pebbles like those of the Boston region were observed at a number of localities. The scarcity of these types suggests that they are the few pieces that travelled 25 to 60 miles from their sources and became mixed with far more abundant stones from nearer sources. Varieties of rock like the

TABLE 3.—*Stone-count data*

Composition (in per cent) of pebbles in the Mashpee pitted plain on Cape Cod

Locality*	T-105	G-95	G-59	G-47	G-12	G-40	T-35	G-61	G-81	G-65	T-79	G-35	Averages: Mashpee pitted plain	Averages: Buzzards Bay moraine	Averages: Sandwich moraine	Averages: Area of overlap of moraines
Granite (including gneissoid granite, aplite, pegmatite)	61	59	71	62	67	70	71	76	75	47	68	72	67	66	45	60
Quartz-feldspar gneiss	5	2	0	0	0	0	0	0	0	1	0	2	1	1	3	0
Diorite and gabbro (including biotite gneiss)	3	3	2	6	6	2	0	0	1	5	1	2	2.5	3	3.5	2
Basalt (including basalt porphyry)	7	4	1	5	0	6	2	2	1	15	3	2	4	3	9.5	5
Other volcanics (mostly rhyolite and pyroclastics)	9	17	6	9	8	4	10	7	11	11	5	6	8.5	4	13.5	11
Schist (including greenstone)	5	2	3	2	7	2	0	1	0	3	7	2	3	3	5	4.5
Slate (including argillite)	2	2	2	1	0	1	0	0	1	0	0	0	1	t	4	2
Quartzite	5	5	7	8	2	6	7	7	5	7	9	7	6	11	7.5	9
Vein quartz	3	7	7	6	9	9	9	6	5	9	6	6	7	8	7	6.5
Arkose and sandstone	0	1	0	0	0	0	0	0	0	1	0	0	t	0	1	t
Conglomerate	0	0	0	0	0	0	0	1	1	0	0	0	t	t	0	0
Miscellaneous	1	0	1	2	1	0	1	0	1	1	1	1	1	1	1	1
	East to West						North		to		South					

* Localities designated T represent collections by Thiesmeyer; G, Goldthwait.

Dedham granodiorite, which make up a large part of every collection of pebbles (*See* Table 3), probably came from buried ledges of bedrock only a few miles away, although no outcrops are visible within 15 miles of the plain. Regardless of the distance they may have travelled, the only plausible sources for these types of rock are north or west of Cape Cod. Southward or eastward transportation by moving ice and its associated melt-water is implied by their presence in the deposits described here.

RELATIONS TO ICE FRONTS

As outlined in preceding paragraphs the Buzzards Bay and Sandwich moraines were deposited by readvance of the ice after construction of the Mashpee pitted plain and overlap the stream-washed sand and gravel at its western and northern margins. The gravel underlying the till of those moraines is in all respects like that of the pitted plain, but its surface is 50 to 100 feet lower. As may be seen along the western side of the Buzzards Bay moraine, however, the surface of the bedded deposits beneath the till is not smooth like much of the plain, and in places large angular blocks are embedded. Therefore, it is believed that these portions of the plain, now buried under the moraine, were originally at the margin of the great fan. There the outwash was deposited on and around irregular masses of ice at the ragged edge of the ice sheet. Shifting currents, such as would result from movements of water over this rough and changing surface, deposited layers sloping at unsystematic angles. The irregularity of the surface increased as the buried ice melted. From the ice itself angular blocks fell and slid into the gravel at some places.

The water that built the pitted plain carried fragments of rock like those found in both the Sandwich moraine on the north and the Buzzards Bay moraine on the west. Counts of the number of different kinds of rocks among 400 pebbles selected at random at each of the 12 locations bear out the more casual observations made at innumerable pits. (*See* Table 3.) Only one stone count, from west of Peters Pond, is closely similar to counts in the Sandwich moraine, and only two stone counts—one west of Massachusetts National Guard Military Reservation (Camp Edwards) on Turpentine Road, the other east of Shallow Pond on Mill Road—are much like counts in the Buzzards Bay moraine. The average of all pebble counts from the plain shows a proportion of granites, diorites, and basalts like that of the Buzzards Bay moraine, whereas the proportion of other volcanic and quartzite pebbles is like that of the Sandwich moraine. As the types of rock contained in the two moraines are significantly different, it would appear that the water that bore gravel to the pitted plain gathered material from sources within both lobes of the ice sheet. During transportation these pebbles were fairly well commingled.

The uniform slope of the plain toward the south and southeast, away from the apex but parallel to each moraine front, shows that the water entered the plain somewhere near or at the present apex, rather than here and there along the line of the moraines. At only two localities (just southeast of Signal Hill in Bourne and just northwest of Falmouth) does the plain show a marked slope away from the bordering moraine. In Bourne this deposit is made of sand and is shaped like a small fan. In Falmouth it is made of coarse gravel already described in the discussion of the Lawrence pit. Presumably, water flowing through breaks in the ridge on the east side of the Buzzards Bay moraine or directly from the face of the Buzzards Bay glacial lobe superimposed fans on top of the main plain at these two localities.

At two other places constructional gravel deposits rise above the general surface of the pitted plain. The more conspicuous one—Falmouth Heights, on the shore of Vineyard Sound—is elliptical, and its top is 35 feet above the adjacent plain. On the seaward side a 40-foot wave-cut cliff consists entirely of cross-bedded sand and gravel. In the upper layers of the gravel are many ventifacts. This gravel must have been deposited as outwash from the ice prior to the building of the pitted plain, for its altitude implies ice or other gravel immediately surrounding the present site to enable streams to deposit at such a high level. This deposition occurred before any other event registered by the surface deposits on this end of Cape Cod. Presumably it occurred while the ice front was retreating from Martha's Vineyard toward the line of the Buzzards Bay and Sandwich moraines.

Originally the deposit may have been far more extensive than it is now. If so, most of it must have been removed by the braided streams that built the outwash fan around it and by ocean waves now active at present sea level. It is more likely, however, that this body of gravel never extended very much farther toward the north than it does today and that it represents the apex of an outwash fan built southward from the ice front when the latter was temporarily standing close by the Heights. Rather rapid recession of the ice then removed the support from the north side of the fan, and subsequent outwash deposits were never built to so great an altitude.

The other outwash deposit above the pitted plain is a discontinuous, broad, low ridge trending southeast from the edge of the Sandwich moraine, north of the Cape Cod Airport, out into the plain as far as the Barnstable-Falmouth road. At its northwest end, this ridge blends into the moraine front; at its southeast end, it is half a mile south of the moraine. Its undulating crest rises 10 to 25 feet above the adjacent plain, and it consists of well-bedded and cross-bedded sandy gravel without boulders. The gentle south slope and slightly steeper north side of the ridge suggest that it, too, was deposited by glacially fed streams when the ice edge stood close to or along its northern margin.

TILL CLUMPS

Masses of unmodified till are not ordinarily found in glacial outwash at any great distance from the moraines with which the outwash is associated. Irregular clumps of unmistakable till were observed, however, at six localities in the Mashpee pitted plain, far out on the glaciofluvial fan. Five of these occurrences were in pits excavated for sand or gravel: (1) immediately west and (2) a mile north of camp headquarters in the Massachusetts National Guard Military Reservation (Camp Edwards); (3) in the northern part of Bear Hollow Farm south of Snake Pond; (4) a mile west of Bear Hollow Farm; and (5) south of the cranberry bogs in Teaticket. The sixth was in a road cut north of Pimlico Pond. Similar till clumps were later noted at a few localities in the pitted plains of the Sagamore and Wareham quadrangles, west of the Cape Cod canal.

The individual masses of till range from a foot or two in length and thickness to more than 10 feet in longest dimension. They are generally more or less rounded or oval in shape although some have irregular or angular bounding surfaces. Commonly there are several clumps within a few yards of each other at any one locality, each entirely surrounded by sand and gravel (Pl. 2, fig. 2). Each locality is at the rim of a kettle hole. No till was found in the scores of gravel pits and road cuts on the smooth parts of the plain or in the many, small, sand pits near the cranberry bogs that have been developed in the furrows at a distance from the kettle holes. Evidently the till was formerly associated with the ice blocks responsible for the kettle holes near which it is now found.

The rock types represented by the pebbles and angular fragments embedded in the sand and clay of the till clumps are like those of the till in the Buzzards Bay moraine, both with regard to variety and relative abundance of the different components. Ordinarily the compactness and coherence of the till are such that the till clumps stand out on the faces of the artificial cuts after exposure to weathering for only a few days or weeks, during which time the surrounding sand and gravel is loosened and blows away or slumps toward the foot of the cut. The surface of each till clump is commonly stained to a depth of 2 to 8 inches with bright red iron oxides, and streaks of the same color penetrate nearly or quite to the center of some of the clumps. Moreover, the sand and gravel immediately adjacent to the till are ordinarily oxidized similarly so that the same red color appears as an aureole surrounding the till at several localities.

All the clumps that have been observed are in the uppermost 6 to 10 feet of the gravel forming the plain, and some extend upward practically to the surface. Where overlain by sand and gravel the covering is ordinarily so thin that in contrast to the subjacent material the bedding has been obscured or destroyed by frost action, the penetration of roots, and the activities of animals. The filling between clumps is relatively loose, non-bedded, or irregularly bedded sand and fine gravel. The uppermost beds of gravel beneath the clumps are wrinkled into wavelike crests in the spaces 1 to 4 feet wide between the bases of adjacent clumps. At the Teaticket locality, the long till clumps stand on end. Near Bear Hollow Farm the long axes of several clumps are all inclined in the same direction. At the National Guard locality, where the till masses lie more or less horizontally, their ends are twisted into a scroll-like form. Where the cuts were deep enough to expose several feet of thickness of gravel below the till, the gravel layers appeared quite undisturbed at depths greater than 2 or 3 feet beneath the base of the clumps.

Evidently the till masses came into place very late in the building of the plain, and considerable pressure was exerted upon at least some of them as they assumed their present positions. Such pressure may have resulted from contraction and expansion of the ice near the top of a stranded ice block exposed to rapidly fluctuating air temperatures. Till dropped directly on the gravel floor beneath the edge of a melting block of ice may have thus been squeezed into place. Moreover, the streams of meltwater on the surrounding plain would occasionally undercut the sides of the dwindling remnants of the glacial ice and cause chunks of it to fall or slide downward and outward, thus shoving the till vigorously into the gravel to produce the wrinkles in the subjacent beds.

SUBAERIAL FURROWS

The surface of the Mashpee pitted plain is not only pock-marked by kettle holes, but it is also scored by long furrows that head about midway up the plain and extend to its outer edge. The longest measures about 10 miles. The lower parts of the larger furrows are now drained by small streams, such as Coonamessett River and Childs River, but these are not actively enlarging the furrows today. The lower end of each is flooded by the sea and forms a long, narrow bay, such as Great Pond, Green Pond, Bowens Pond, and Eel Pond (Fig. 3).

Erosion rather than deposition is responsible for these furrows, since they are actually cut into the sand and gravel layers of the plain, as indicated by the truncated edges of gravel beds seen in exposures on the walls of furrows. Two of the best exposures available in 1939 were in road cuts half a mile southeast of Jenkins Pond (Pl. 2, fig. 3) and a quarter of a mile southeast of Round Pond. Furthermore, the furrows have the form of stream-cut valleys. Each heads in a narrow, shallow, V-shaped cut. Down valley, most of them increase to approximately 1000 feet in width and 25 feet in depth. Their floors are broad and smooth. Short, V-shaped tributaries enter the main valleys and give each system a stunted dendritic pattern like the veins of a narrow leaf.

The close spacing of the furrows and subparallel orientation of all their main courses are remarkable. Some adjacent furrows are so close where they enter Nantucket and Vineyard Sounds that the intervening flat-topped remnants are rarely as much as half a mile broad. All are oriented within 20° of south. Those near the western edge of the plain are nearly parallel to the Buzzards Bay moraine. Such an arrangement of valleys is not like the dendritic "crows-foot" pattern formed in the normal processes of stream erosion, and it, therefore, requires some special explanation.

As many furrows are not now occupied by streams, present-day erosion is totally inadequate to account for these forms. As the headwaters are not directly connected with areas of moraine, and as the pattern of the furrows is not braided and therefore is not indicative of overloaded streams, the furrows were not cut by the glacial melt water that constructed the plain. It is possible that, after the plain was essentially complete, the extra supply of water from ice melting west and north of the moraines saturated the loose gravels. This ground water might have reached the surface in springs part way down the fan and flowed on the surface from there to the sea. A second possibility is that there was a period of barren, treeless climate, caused either by drouth or cold, during which ephemeral rains cut shallow arroyos in the gravels. It is also possible that the condition of the ground, frozen while the ice sheet still covered the interior of New England, made all save surface gravels impermeable to water and induced shallow cutting in closely spaced channels.

PERSISTENCE OF BURIED ICE

The duration of all these events which contributed to the shaping of the present Mashpee pitted plain is very impressive. Following the isolation of large masses of ice from the wasting ice sheet, braided streams of melt water deposited sand and gravel around and over most of the outlying ice masses. The period required for the construction of such a large plain, which now contains more than 2 cubic miles of gravel, must have been many tens and perhaps hundreds of years.

The surface materials of the plain include many thousands of stones which are plainly cut, grooved, and faceted by the wind. These may be found singly or in quantity in nearly half of the pits now excavated. The deep etching of these ventifacts and their abundance impress all who see them with the vigor and duration of wind action that must have taken place. No comparable wind action is effective on the Cape today. Many of these ancient ventifacts are actually in the beds of gravel near the surface of the plain, but they could not have rolled far in the streams that built that portion of the plain or the delicate polish and edges made by the wind would have been obliterated. It is necessary, then, to assume that the deposition of the upper layers of the plain alone required many tens and perhaps hundreds of years during which wind action was very intense.

Buried ice persisted under the sand and gravel of the plain all the time that the wind was thus active, for none of the sand or silt that must have supplied abrasive tools for the sandblast fills the kettle holes. Furthermore, the furrows were cut before the ice entirely disappeared. Furrows are found to enter and exit from groups of kettle holes with no change of direction. There is no sign of drainage from one kettle hole to another. Dozens of small pits occur on the floors of furrows, and large pits may interrupt and separate a whole headward section of furrow from its lower flooded end at the sea. Had these low kettle-hole depressions existed when the furrows were cut, they would certainly have been filled with stream gravel to the level of the channel floor. As this is not true, the ice must have remained after the plain was built until the furrows had been cut.

Buried ice remained while the adjacent moraines were superimposed upon the edges of the plain. Thin active ice moved forward over this gravel and deposited the till of the Buzzards Bay and Sandwich moraines without destroying the layers of gravel beneath. Masses of stagnant ice caught under the first outwash gravel remained until the active ice overrode them, inasmuch as certain large kettle holes already described extend from the exposed surface of the fan into the moraine-covered area. The ice finally melted away after both the gravel of the fan and the till of the moraine had been deposited.

DROWNED SHORE LINE

The final event affecting the present aspect of the Mashpee pitted plain was drowning by the sea. At no time since completion of the plain has sea level been higher on this portion of Cape Cod than it is today, for no sand bars were found above the modern shore line. These would certainly be present had wave action once been higher on such a sandy shore. Moreover, none of the kettle holes at the lower edge of the plain are filled by wave wash. On the other hand, submerged cedar stumps have been dredged from below low-tide level at Witchmere Harbor in Harwichport, 15 miles east of this pitted plain (Chute, 1939, p. 25), and at Centerville (Sawyer, 1932) near its eastern edge. The lower end of each furrow is obviously drowned; no process of shore erosion or construction could produce such a deeply indented shore line. Following the rise of the sea to its present stand with respect to this part of Cape Cod, sand bars were built across many of the bays and estuaries.

Low wave-cut cliffs facing narrow sea beaches have been eroded along the low margin of the plain. If the original gentle surface of the plain be projected beyond the present cliffs until it touches sea level, it appears that the plain once extended at least a quarter and perhaps half a mile farther seaward. Waves and shore currents have beaten the shore line back by this amount. At distances greater than half a mile off the Falmouth shore, or one mile off the shore at Cotuit, the Coast and Geodetic Survey charts (no. 1209 especially) show a submarine slope far too steep to be the original surface of the pitted plain. If the pitted plain formerly extended beyond that line, recent excavation by currents, or wave cutting at some lower stand of the sea, has modified the bottom contour too much to permit reconstruction of the original outer margin of the plain.

PART III. LATE GLACIAL VENTIFACTS

OCCURRENCE AND DISTRIBUTION

During these investigations on western Cape Cod thousands of wind-worn stones or ventifacts (Bryan, 1931) were found embedded in the various till masses and glacio-fluvial deposits. Reconnaissance at other widely scattered localities on or near the Cape—Orleans, Brewster, Wellfleet, Chatham, and Plymouth—disclosed a similar remarkable abundance of ventifacts in the glacial and postglacial deposits. This confirmed reports of several writers (Woodworth, 1894; Davis, 1895; Bryan, 1932) that aeolian action was prevalent in this part of New England during or since the retreat of the last Pleistocene ice sheet. More than 2000 of these wind-sculptured stones were collected and studied, not only because they exhibited great variety in shape and in type of abrasion but also for the information they might provide concerning the age of the deposits enclosing them.

The ventifacts were found in various relations to their surroundings. Their mode of occurrence may be described under four categories.

(1) Some were lying on the surface. Many of the stones thus noted

appear to have reached their present position largely through overturn of the upper few feet of material by frost action, though some may represent residual accumulations from slope wash. Ventifacts are also fairly common on beaches derived from shoreline erosion of till or gravel bluffs, as at Nobska Point (Woods Hole quadrangle), Wings Neck (Onset quadrangle), Falmouth Heights (Falmouth quadrangle), and Sagamore Heights (Sagamore quadrangle).

(2) Others occur in the upper soil zone and in the disturbed and weathered zone, or "warp." The warp is a thin mantle over undisturbed till or over gravel layers. Where the thickness of the disturbed material was observed, it generally approximated 2½ to 3 feet and rarely was as great as 4 feet. There is good reason to believe, moreover, that very little material has been removed by slope wash or wind action at many localities. The depth of frost action was not very great, therefore, despite the periglacial climate that prevailed during recession of the ice.

(3) Some of the ventifacts are at the top of the glacio-fluvial gravels and beneath well-sorted sand, several feet thick. Although this mode of occurrence is uncommon, it is well exemplified in a gravel pit overlooking Whitcomb's Bog (Pocasset quadrangle) (Pl. 3, fig. 1). There, scores of ventifacts of many sizes, including small boulders, lie closely packed along a single horizon, apparently undisturbed since they were cut. Only the top and sides of each stone display signs of wind work.

(4) Many ventifacts are embedded in undisturbed till and gravel. Specimens were collected from layers of well-bedded gravel, 2 to 8 feet below the disturbed zone at Falmouth Heights, 10 feet below at Whitcomb's Bog, 8 to 20 feet below at Nameloc Heights (Sagamore quadrangle), and from till, 2 to 8 feet below the "warp" at Woods Hole, at Gunning Point (Woods Hole quadrangle) and at Wings Neck. The ventifacts embedded in till have, of course, a random orientation, and many show signs of abrasion subsequent to wind-cutting. Presumably, they were picked up during minor readvance of the ice which deposited the till that enclosed them. Some of the stones embedded in gravel are scattered at random and with varied orientations of the wind-smoothed surfaces. Many of them show varying degrees of abrasion subsequent to windcutting, which must in part at least be a function of the distance they were carried by the melt waters that brought them to their present position. Other ventifacts embedded in gravel occur in groups along certain horizons and appear to have been simply buried with little if any disturbance by the superjacent layers of gravel or sand.

Wind-cut stones can be found associated with glacial and glacio-fluvial deposits almost everywhere on Cape Cod and in adjacent areas of New England (Shaler, 1889, Pl. X; Woodworth, 1894), yet they are

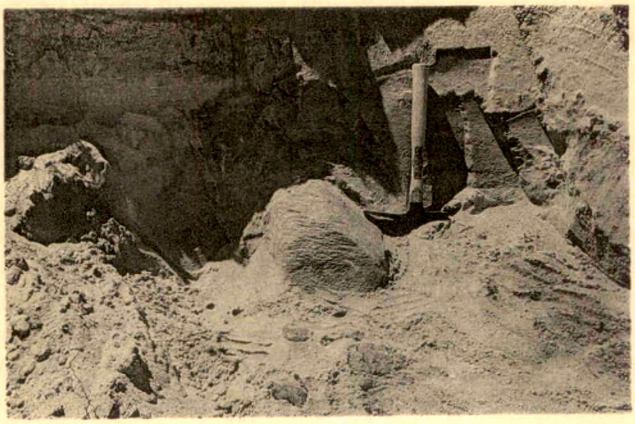
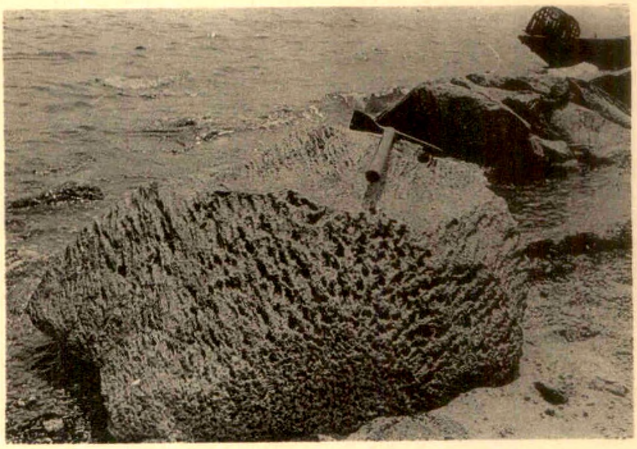


Figure 1. Granite Gneiss Boulder with Fluting Transecting Foliation

Boulder is at top of bedded gravel beneath about 6 feet of well-sorted sand, exposed in pit at side of Whitcomb's Bog.

Figure 2. Granite Block on Shore of Cape Cod Bay

Near Stage Point, 3 miles north of Sagamore quadrangle. Pits and grooves radiate from windward end in foreground. They are 1½ inches deep where impact of wind-driven sand was direct and grade to shallow flutings on sides of stone.



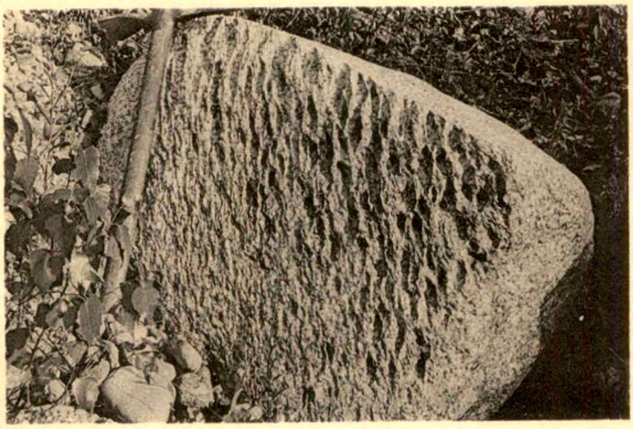


Figure 3. Deeply Sculptured Granite Boulder

Grooves and pits are 1 to 2 inches deep.

VENTIFACTS PHOTOGRAPHED IN THE FIELD

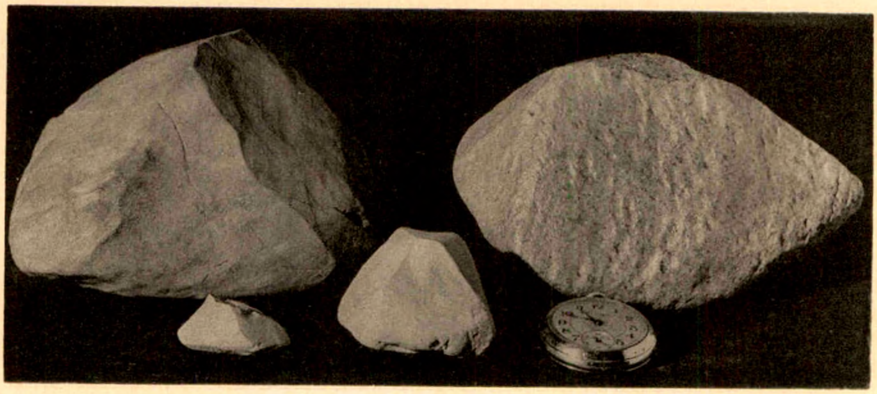


FIGURE 1. VENTIFACTS TYPICAL OF THOSE DISPLAYING PYRAMIDAL FORM
Quartzite at left rear has 16 interfacet edges, only a few of which are shown. Small ventifacts in the foreground are aplites; the one at right rear is a felsite porphyry.

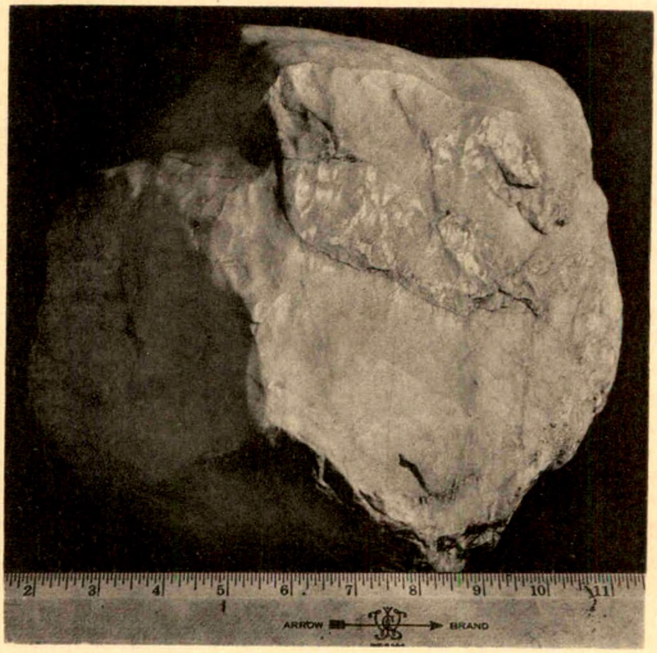


FIGURE 2. HIGHLY POLISHED SURFACE OF QUARTZITE VENTIFACT
Light reflected from conchoidal surfaces below sharp interfacet edge near top shows patchy preservation of cellophanelike polish.

PYRAMIDAL VENTIFACTS AND WIND POLISH

very unevenly distributed. One pit may yield hundreds, whereas pits near-by seem totally devoid of them. Even within a single exposure there is no uniformity in occurrence. Ordinarily they are sparsely distributed in both till and gravel, but patches of till a few feet in diameter and certain parts of single gravel layers may yield hundreds of specimens. In general, the ventifacts seem to be more common in the upper 4 to 8 feet of the deposits.

ROCK TYPES REPRESENTED

The glacial deposits of Cape Cod were derived from highly varied lithologic units comprising the bedrock in parts of eastern and northern New England. It was not surprising, therefore, to find more than 30 distinct rock types among the wind-worn stones. The rock species which were separately recognized in the writers' collection are listed below; opinions, based on lithologic comparisons, concerning the probable derivation of some of them are indicated. Such opinions are shared with several geologists conversant with the areal geology of New England who have visited these areas or examined specimens from them.

The lithologic types distinguished among Cape Cod ventifacts are:

Vein quartz (pure milky, blue brecciated, chloritic, jasperized, and colorless varieties)
Pegmatite and graphic granite (pink, yellow, and gray varieties)
Quartz-pink-feldspar gneiss
Aplite
Fine-grained granite and granite gneiss (probably fine-grained facies of the coarser types)
Medium to coarse granite and granite porphyry (massive to gneissose)
    Dedham granodiorite type
    Quincy granite type
    Types common in Maine ?
    Milford granite type ?
Diorite and gabbro-diorite
    Salem type
Diabase
    Medford type
Fine-grained mafic rocks (massive, porphyritic, and amygdaloidal)
Volcanic rocks of rhyolitic to intermediate composition
    Agglomerates and breccias
        Types found at Nantasket
    Felsite porphyries
      and
    Flow rhyolites
        Lynn volcanic types and type associated with Quincy granite
    Amygdaloidal
Conglomerates
    Roxbury conglomerate types
Quartzites (white, brown, blue-black, red, black and white varieties)

Argillite, phyllite, and slate (gray, green, red, laminated varieties)
    Cambridge slate types
Biotite gneisses
Schists (biotite, chlorite, black and white quartzose laminated)

SIZES AND SHAPES

The wind-cut stones range from pebbles less than an inch in diameter to enormous boulders. Many large granite blocks have more than 20 square feet of wind-etched, fluted, or polished surface. Excellent examples may be seen on the beach at Wings Neck, and at Sagamore Heights. Many others were found in place in glacial till or gravel. The largest single wind-cut surface seen is the top of a huge boulder, known as Sacrifice Rock, in Bournedale, which measures approximately 30 by 20 feet. Its entire upper surface was polished and slightly fluted by wind abrasion.

So great a variety of shapes was found in the collection as a whole and among the stones gathered at single localities that there is no basis for conclusions concerning constancy or variability of winds, such as have been developed by other writers from experimental and field evidence. Eddying currents must have been produced at many places in the wind stream by the larger stones, and smaller ones lying on their leeward sides were thus protected to varying degrees. The shapes produced on these would consequently be far different from those which result when single, isolated specimens are attacked by sand blown across a desert floor or a laboratory table. Furthermore, the study of Cape Cod ventifacts revealed nothing to substantiate Heim's conclusion (1922, p. 349) that the number of facets that will develop on a stone undergoing wind abrasion is determined by the shape of its base. On the other hand it appeared that, in this collection at least, the shape of the original fragment was the most important factor controlling the present shape of the ventifact.

The following general shapes were found among the wind-faceted stones of the Cape, along with many forms intermediate between or representing combinations of these types:

(1) *Einkanter* (single facets bevelling rounded stones)—single, flat or convex facets truncating the curvature of more or less symmetrical, water-worn cobbles. These, however, are uncommon; most of the stones show truncation on more than one side.

(2) *Firstkanter* (ridge-shaped)—two major facets sloping away from each other and forming a ridge; includes shapes resembling spearpoints or arrowheads (Pl. 7, fig. 2).

(3) Pyramidal—three or more facets sloping from a common center (Pl. 4, fig. 1); includes *dreikanter*, *fünfkanter*, and so on. Contrary to

expectations from the impression generally conveyed in textbooks, *dreikanter* are the exception among these wind-cut stones. Less than 20 were found, although many others approach that shape.

(4) Triquetous or brazil-nut—three facets, commonly slightly curved, intersecting at about 60° (Pl. 5, fig. 1).

(5) Polygonal—number of facets variable and probably determined largely by the number of fracture surfaces on the block prior to sandblasting.

(6) Irregular—shape and number of facets very variable; this group may include early stages in the formation of the types listed above.

The length of time different facets on the same specimen were exposed to wind erosion, as judged by the depth of pits and flutings and the amount of material removed, varied considerably. This is more readily explained as a result of rolling of the specimen than as an effect of variable winds.

Even less of a tendency to uniformity in shape was discernible among the smaller stones, though they have been subjected to a more varied attack such as would affect the larger ones in time through continued erosion and rotation. This opens to question the concept advanced by some writers (Schoewe, 1932, p. 125) that the brazil-nut or triquetous shape represents the ultimate form that all ventifacts, regardless of their texture, should acquire in time, given a strong wind of constant direction.

EVIDENCE OF SANDBLASTING

The evidence given below makes inescapable the conclusion that the stones considered in this paper are truly ventifacts, shaped and scoured to varying degrees by wind-borne sand and dust. The features listed have all been ascribed by various writers to aeolian abrasion; they have been observed on ventifacts from modern deserts and beaches; and they are difficult to explain except as results of sandblasting. Consequently, they serve as useful criteria for recognizing ventifacts. All the specimens collected by the authors show one or more of them.

(1) Polished surfaces. Very few of the specimens have the high luster characteristic of ventifacts from modern deserts, but the preservation of patches of such a polish on some (Pl. 4, fig. 2) and of a somewhat duller luster on at least parts of the rest suggests that originally most of the wind-cut surfaces were highly polished. Gentle buffing with a soft cloth may be necessary to remove a film of dust that conceals the gloss. Commonly the polish is preserved only by quartz grains in granitic rocks; but some feldspar grains also have a somewhat duller luster. Quartz and quartzite are more lustrous in general than the feldspathic rocks.

(2) Smooth, greasy surfaces. Except where this characteristic has been destroyed by weathering or abrasion, wind-carved surfaces are smooth and feel distinctly greasy, like talcum powder. This is true not only of the larger faceted and cuspate surfaces but also of the inner parts of grooves, pits, and hollows, and along the windward sides of low projections produced by differential etching.

(3) Wind-smoothed faces and wind-cut facets. These range from less than half an inch square to many square feet. Most are either concave or convex, but a few are essentially flat. The number of such surfaces on any single stone varies considerably. Some of the facets probably represent merely modified fracture surfaces, but those that bevel rounded cobbles and those that form pyramidal shapes were clearly produced by wind abrasion (Pl. 3, fig. 3; Pl. 4, fig. 1).

(4) Sharp interfacet edges. These range from knifelike sharpness on some quartzite cobbles to rounded and pitted ridges on water-worn specimens (Pl. 4, fig. 1; Pl. 5, fig. 2). Edges formed at the junction of fluted facets are serrate (Pl. 6, fig. 1).

(5) Differential etching. Veins, phenocrysts, inclusions, fragments in volcanic rocks, quartz blebs, laminae, pebbles in conglomerate, and amygdules, commonly stand in relief on windswept surfaces or are etched below the general level of their surrounding media (Pl. 7, fig. 1). In either case a well-defined luster, a smooth and greasy feel, and occasional small-scale flutings show that they were produced by differential etching in a sandblast rather than by differential solution during weathering. Some differential solution may have preceded the wind action on a few of the stones so that the irregularities were accentuated by the wind, but aeolian abrasion alone could have produced their present shapes. Several ventifacts with such etching have bizarre shapes, representing small-scale examples of the peculiar erosion forms seen on landscapes in arid and semi-arid regions (Pl. 6, fig. 2).

(6) Fluted surfaces. Differential resistance of the various constituent minerals in granular igneous rocks expresses itself in groups of discontinuous, subparallel, concave grooves or flutings. These form more or less denticulated surfaces that range from almost microscopic indentations to 1½ inches in depth (Pl. 3, figs. 2, 3; Pl. 6, fig. 1). They were carved by wind moving essentially parallel to the rock surface on which they occur. Their orientation is independent, therefore, of pre-existing structure in the rock and is parallel to the wind motion. They may cut diagonally across gneissic banding, but on some specimens, or on parts of one specimen, they may fortuitously follow it.

On the coarser-textured ventifacts, fluting is universally present. It may be a subordinate feature of well-formed, wind-carved facets (Pl. 5,

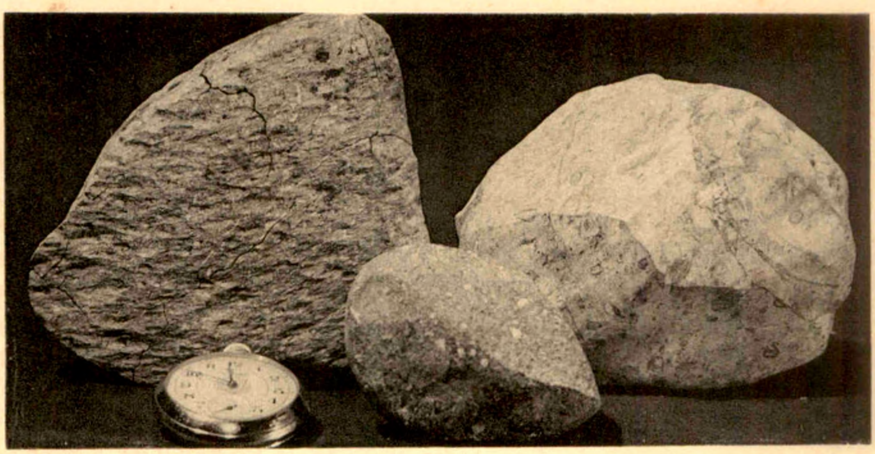


FIGURE 1. VENTIFACTS SCULPTURED FROM VOLCANIC ROCKS
Shallow flutings on basalt porphyry at left were controlled by etching out of phenocrysts. Recrystallized rhyolite tuff at right is denser, and fragments were as resistant as matrix. Phenocrysts in basalt porphyry in middle exerted little influence on ultimate triquetous form.

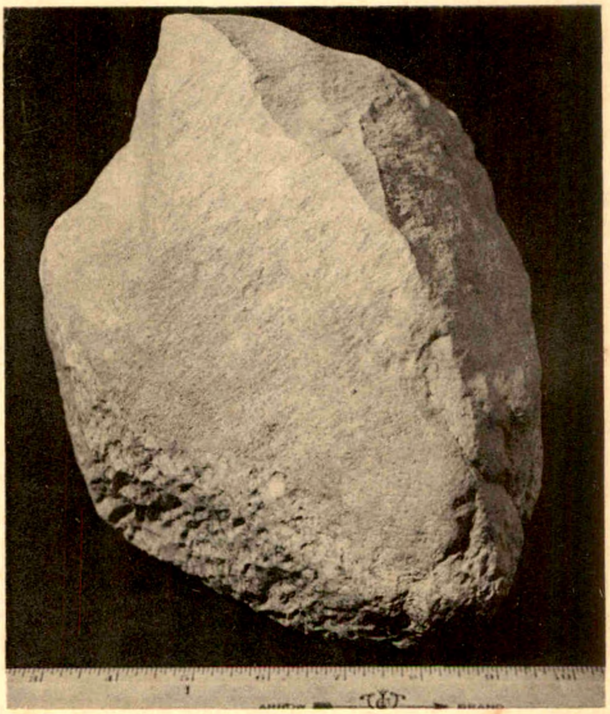


FIGURE 2. VENTIFACT CARVED PARTLY FROM GRANITE AND PARTLY FROM DIKE ROCK
Sharp edges and smooth facets are typical of fine-textured rocks, irregular pits and flutings of coarse texture.

INFLUENCE OF ROCK TEXTURE ON FORMATION OF VENTIFACTS

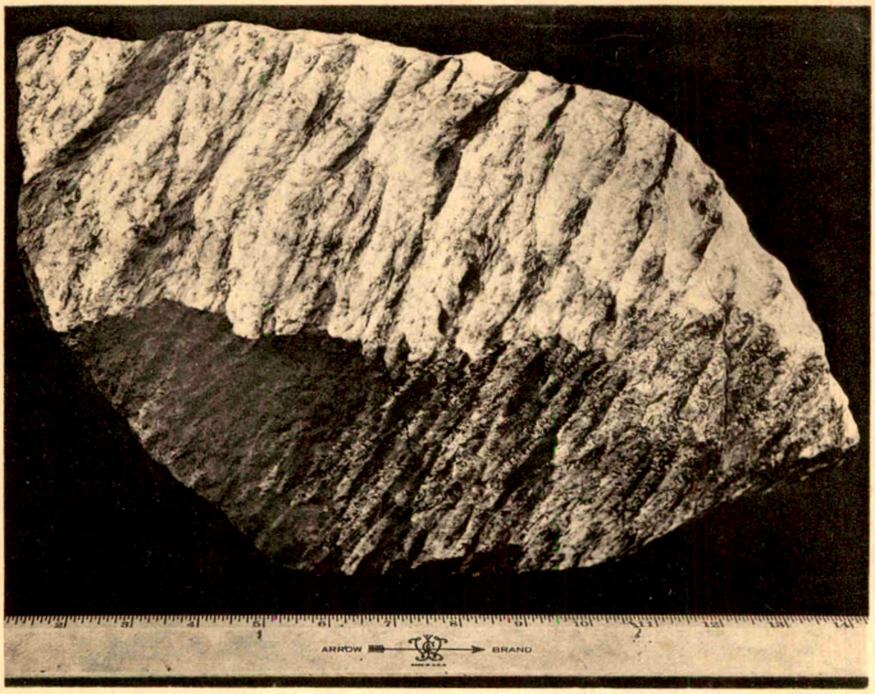


FIGURE 1. DEEPLY FLUTED GRANITE GNEISS
Each side is a fluted facet; orientation of grooves indicates rotation of stone on long axis.

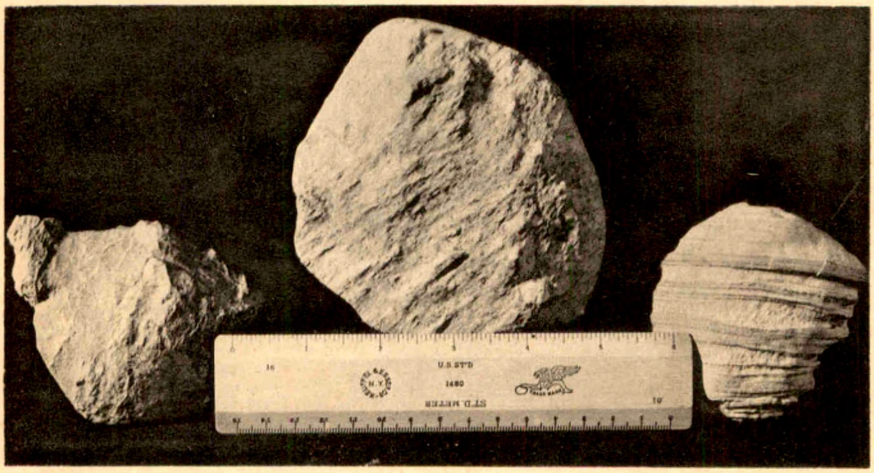


FIGURE 2. DIFFERENTIAL ETCHING OF THREE LITHOLOGIC TYPES
Greenstone at left, porphyry in middle, and fine-grained gneiss at right. Trails on leeward side of projecting phenocrysts are notable on surface of porphyry.

GROOVED AND ETCHED VENTIFACTS

fig. 1) or it may occur on rather flat surfaces which represent modified fractures. Fine-textured rocks, such as volcanic porphyries or gneisses, containing elements of different hardness, may have fluted facets; but homogeneous, hard quartzite, aplite, and vein-quartz pebbles do not ordinarily show fluting.

The arrangement and character of the flutings may furnish excellent evidence concerning the wind direction and other conditions under which a certain ventifact was produced. For example, on some specimens the fluting is diagonal to interfacet edges (Pl. 3, fig. 2). This condition eliminates the possibility that such ventifacts originated when opposing facets were developed simultaneously by a sandblast directed essentially parallel to the interfacet edge. It also renders inapplicable to these particular stones the observation of Woodworth (1894, p. 69) that sloping facets may be produced in a position more or less normal to the wind. Again, the orientation of flutings on adjacent facets may show not only that a stone was dislodged and rotated into a new position before each succeeding facet was produced, but it also may indicate the directions and axes of such rotation (Pl. 6, fig. 1).

(7) Pits and cuspate hollows. Shallow to deep pits may occur singly or in groups on both smooth and fluted facets, representing the etching out of phenocrysts or other less resistant portions of the stone. The size, shape, and distribution of the hollows are controlled by patterns of nonresistant elements in the rock (Pl. 5, fig. 1).

Several fluted and faceted granite boulders show that they split the wind. On their windward ends, where the surface of impact was nearly normal to the sandblast, deep pits were formed by etching out of feldspars. The pits deepen to leeward, and some have overhanging, rooflike projections of more resistant quartz or feldspar at their leeward terminations. On surfaces curving upward to facets on the top and sides of the boulders (where the sandblast moved more nearly parallel to the erosion surface) these pits become progressively shallower and increasingly elongated parallel to the wind until finally they grade into ordinary flutings on the flatter parts of the facets. The pits, hollows, and flutings display a definite tendency to radiate from the windward ends of the boulders (Pl. 3, fig. 2).

One specimen of dense greenstone has peculiar cuspate hollows on its windward side directly beneath a projecting ridge of quartzose vein material that forms its upper surface. These cusps also deepen to leeward more rapidly than those described above so that they are essentially inclined and cup-shaped. Lateral coalescence of several of them formed a scalloped edge in front of the quartz overhang. Smoothness and polish of the bottoms of the cusps, of the windward face of the rock, and of the projecting ledge show that this phenomenon is properly to be attributed to aeolian action rather than to solution.

Quartz and quartzite ventifacts commonly have broad, shallow cusps on well-developed facets. These are not of the same origin as the pits described above but were probably produced by modification of the curvatures normally formed by conchoidal fracture.

(8) Effects on small stones. Almost all the characteristics of wind abrasion found on the larger ventifacts are duplicated on those which range from 2 inches to less than half an inch in diameter (Pl. 7). These smaller ventifacts show likewise that their texture and mineral content were important in controlling the development of fluting, pits, and facets and that rotation during wind-cutting determined the number and perfection of their facets. These small-scale "replicas" were, however, affected on all sides more commonly than their larger counterparts, doubtless because they became dislodged and were rotated more readily and more frequently.

ABRASION DURING SUBSEQUENT TRANSPORTATION

Many of the ventifacts were modified to varying degrees by abrasion during glacial or glacio-fluvial transportation. This is seen in destruction of the polish, rounding of the interfacet edges, lowering of the ridges between adjacent grooves or flutings, and replacement of the smooth, greasy surfaces by lusterless, minutely pitted ones that have a harsh and gritty feel. Several stones were found whose only remaining evidence of wind-cutting consisted of patches of unmistakable fluting, glossy and smooth in their deeper portions. Hundreds of ventifacts, on which facets and other indications of wind-scouring were almost obliterated by weathering and abrasion, were seen but not collected. Still other hundreds were no doubt overlooked because they resemble too closely the normal rounded to subangular stones of the glacial deposits. This is probably true also in other glaciated or driftless areas where the existence of ventifacts associated with the Pleistocene deposits has not been reported. There is little evidence of abrasion of this type on the abundant ventifacts in the upper 8 to 10 feet of the gravel deposits. They could not have been carried far, therefore, after the wind had shaped them. The lack of glacial striae on the modified facets of ventifacts embedded in till is a further indication of short transport from the places where sandblasting occurred to present positions in the till.

WIND-CUT SURFACES TRANSECTING WEATHERED RIMS

General statement.—On a few ventifacts of differing lithologic types (one granite, two greenstone volcanics, one quartzite) the facets were developed across narrow limonitized and hydrated rims. These stones

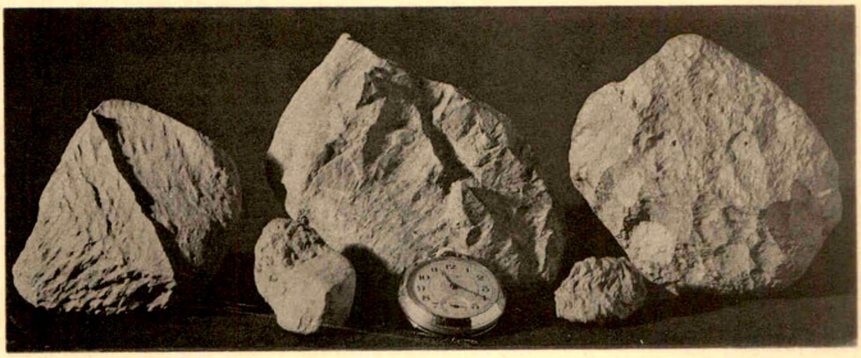


FIGURE 1. DIFFERENTIAL ETCHING BY SANDBLAST
Rocks are greenstones and volcanic breccias.

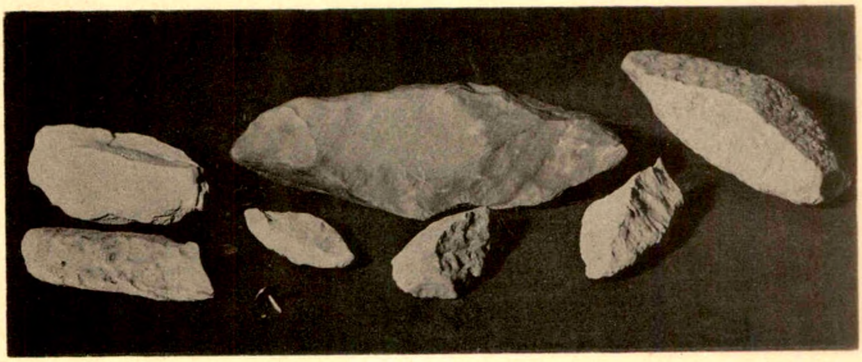


FIGURE 2. SPEAR-SHAPED VENTIFACTS
Quartzite in middle rear has cusps on modified surfaces of conchoidal fracture; other specimens are granites and felsites.

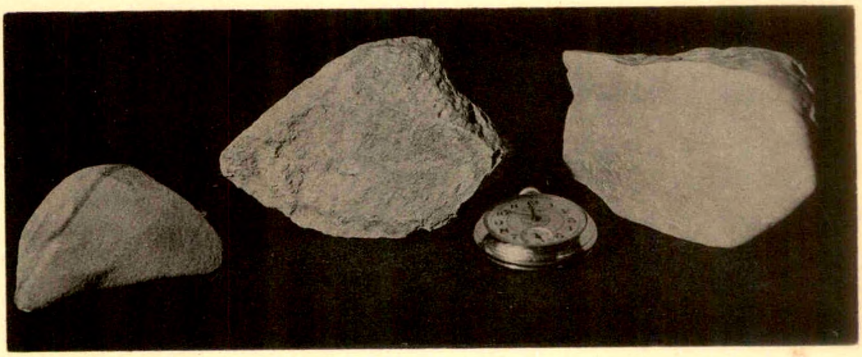


FIGURE 3. WIND-CUT FACETS TRUNCATING WEATHERED RIMS
Limonitized selvage on felsite porphyry in middle is cut by sandblasted surface at upper right; lower front face is fracture broken to show complete weathered rim. Other specimens are greenstones with wind-cut facets facing camera; darker zones are the result of weathering before exposure to sandblast.

VENTIFACTS OF VARIOUS SHAPES
Some showing weathering before sand-blasting.

were broken through with a hammer to prove that the rims were truly formed by centripetal penetration of weathering agencies rather than by superficial staining while the specimens lay embedded in the soil. The facets were cut deeply enough to expose the fresher cores of these rocks so that now the brown rims form a border selvage on the facets (Pl. 7, fig. 3). On one specimen the formation of the facet removed a whole side of the rounded stone so that no such border remains on that side. Another greenstone cobble was rolled and so deeply faceted on all sides that only a few patches of its weathered rim have survived.

The facets on most of the ventifacts do not show similar staining or comparable weathering. In fact, some have been so little affected by weathering since they were wind-cut that they have retained a greasy, lustrous surface. In a rock as susceptible to weathering as greenstone this can only mean that little time has elapsed since the facets were carved, unless an appeal be made to extraordinarily perfect protection from weathering processes.

Date of formation of weathered rims.—No certain way of dating these uncommon rims has yet been found. Since the weathered rims obviously were formed prior to the facets that cross them, and the facets were carved during recession of glaciers believed to represent the Wisconsin stage on Cape Cod, the weathering must be assigned to a pre-Wisconsin substage of glacial time. It does not seem likely that cobbles weathered during pre-Illinoian time could have survived Illinoian transportation, Sangamon weathering, and Wisconsin transportation. It is possible, however, that the present weathered rims are merely the still fairly compact inner parts of much thicker shells of decomposition which were largely removed during these stages. Too few similarly weathered rims were seen on boulders and cobbles that escaped wind-cutting to support the suggestion that the rims were formed during the short interval just preceding the wind-cutting—that is, between the time the stones were washed out of receding ice and the beginning of wind wear. Determination of the precise age of this weathering is a subject for further investigation. Nevertheless, the finding of such partly weathered stones in glacial drift among the myriads of almost fresh ones increases the necessity for taking into account the possibility of preglacial or interglacial weathering of some stones in applying the comparative weathering of constituent stones as a criterion of age in separating drift deposits (MacClintock, 1940).

SIGNIFICANCE OF VENTIFACTS IN DATING DEPOSITS

The widespread occurrence of ventifacts within the till masses and outwash gravels of western Cape Cod indicates that they were sculptured under periglacial conditions during recession and local readvance of the ice. Modifications of the wind-scoured surfaces during transport after wind-cutting could have developed only if aeolian abrasion was in progress during construction of each of the successive glacio-fluvial landforms and prior to the development of each moraine in which ventifacts are embedded. Stones lying on a surface of outwash were sandblasted and then picked up in a readvance of the ice to become incorporated in the till of the resulting push moraine. This sequence of processes was doubtless repeated many times during the recession of the ice from its terminal position.

The ventifacts support and corroborate other lines of evidence that no great length of time has elapsed since the deposits of western Cape Cod were made. The freshness of all the wind-cut surfaces, including those across earlier weathered rims, and of the associated stones, argues strongly against a date earlier than the Wisconsin stage for their formation. The preservation of polish on many ventifacts in the surface zones of the deposits points to the same conclusion. These stones in the disturbed "warp" cannot be assigned to post-Wisconsin wind action because there is so little wind-drifted sand on the surface, even in the bottoms of numerous kettles—no anchored dunes, loess deposits, or patches of aeolian sand capable of forming the many thousands of ventifacts. Stetson (personal communication, 1940) reports abundant frosted and well-sorted sand (probably aeolian) on the bottom of Vineyard Sound, which may represent the missing abrasive. If so, it blew off the present area of the part of the Cape under consideration before the kettle holes were formed, for such hollows could not have escaped at least partial filling had they existed when the ventifact-producing sands swept across this region. These stones must have been shaped, therefore, before all the buried blocks of ice had disappeared. This does not harmonize with the suggestion advanced by some geologists that the ventifacts were formed at the surface in postglacial time and churned back with random orientations into the ground by vigorous frost action. The evidence given above seems to indicate, on the contrary, that both the ventifacts and the deposits that contain them belong to a late Wisconsin stage.

CLIMATIC IMPLICATIONS

Although in abundance and perfection the ventifacts are clear indicators of strong winds blowing down from oscillating margins of late Wisconsin ice, they do not signify an arid climate. The thickness, areal extent, coarseness, and channeled character of outwash gravels on western Cape Cod indicate strong discharge of meltwaters, periodically at least, during the accumulation of such debris. Nevertheless, ablation must also have been an important process in the downwasting and disappearance of the ice because there was apparently little subaerial erosion of pitted outwash plains by meltwaters while recessional moraines north of them were still being formed by active ice. The climate must have been somewhat arid, however, to permit large quantities of sand to be drifted about the area and to prevent the flourishing of vegetation which would have anchored the sand and would probably have caused more pronounced chemical decomposition of the constituents of the deposits than has been observed. That there was only slight to moderate abrasion of many of the ventifacts during glacio-fluvial transport strongly suggests short transfer, such as would occur in sudden flushing of meltwaters, torrential rains, or rapid shifting of channels. The character of subaerial valleys (furrows) on the pitted plains is strikingly similar to that of arroyos carved under semi-arid conditions. The authors conceive, then, of the sculpturing of these ventifacts and disappearance of the ice in a disagreeable, semi-arid climate characterized by sporadic heavy precipitation, strong winds, and low temperatures.

SUMMARY

Study of more than 2000 ventifacts embedded in undisturbed deposits and in disturbed surficial zones of western Cape Cod yields corroborative evidence that the deposits were formed during late Wisconsin time. The climate during recession of the ice and sculpturing of the ventifacts was cold and semi-arid. The present shapes of the wind-cut stones were determined by the shapes of the original fragments, textural and structural features of the individual rock, amount of dislodgment and rotation during sandblasting, duration of abrasion while a stone lay in one position, and degree of protection from direct impact afforded by adjacent stones. Because such a large number of lithologic types is represented among these ventifacts a variety of erosional effects which may be used as criteria of sandblasting were observed. Modification of wind-scoured surfaces during later glacial and glacio-fluvial transportation shows that aeolian abrasion occurred during construction of the various landforms made of outwash and prior to the development of each successive recessional moraine. Lack of an appreciable veneer of wind-drifted sand over the surface, even in the bottoms of kettles, indicates that cutting of the ventifacts was completed before all of the residual buried ice blocks had disappeared.

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ORIGIN AND STRUCTURE OF THE LIVINGSTON IGNEOUS ROCKS, MONTANA

BY WILLARD H. PARSONS

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ABSTRACT

Late Cretaceous andesitic volcanic rocks immediately north of the Beartooth Mountain uplift and included in the Livingston formation originated from many small vents near the headwaters of the Deer creeks. The series is composed of volcanic

breccias and tuffs near the vents and of volcanic conglomerates and sandstones farther away. The latter are interbedded with Judith River to Lance (upper Cretaceous) sediments. Pipes, dikes, and laccoliths of dacite, andesite, and diorite cut the pyroclastics and near-by sediments. Some breccia beds and the underlying sedimentary rocks have been tilted by the intrusion of several laccoliths prior to the Laramide orogeny. Other pyroclastics are later than the laccoliths but pre-Lance. Igneous activity had ceased before the main Beartooth thrusting.

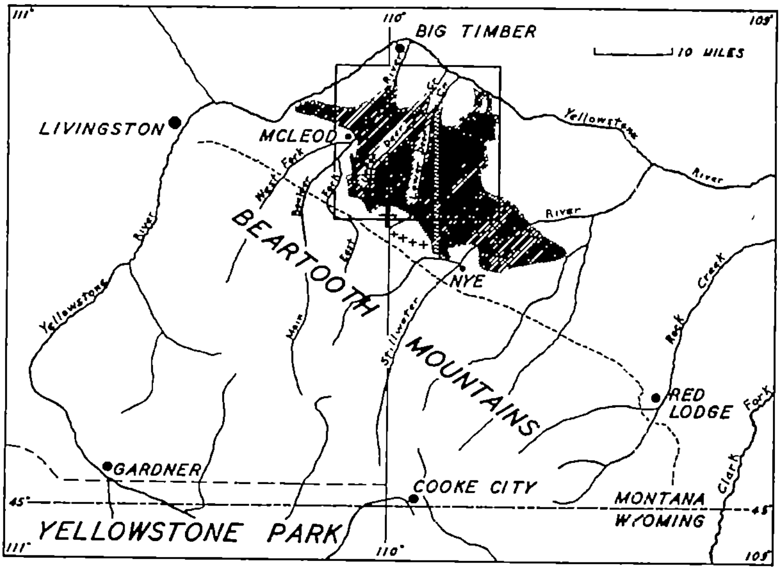

FIGURE 1.—*Sketch map of the Beartooth Mountain area*

Showing location of the Livingston igneous series (diagonal lined area). Broken line is approximate position of the Beartooth Mountain front. Crosses represent laccoliths (larger crosses = larger laccoliths). Small rectangle demarcates area of Figure 2, and area discussed in this paper

INTRODUCTION

PURPOSE AND LOCATION

This study deals with the structure, origin, and age relations of the pyroclastic and intrusive rocks, the former a part of the Livingston formation, in the Deer Creek-McLeod area, Montana. The location of this area is shown in the accompanying sketch map (Fig. 1). These rocks represent one unit of the many late Cretaceous volcanic rocks of western Montana.

ACKNOWLEDGMENTS

Field work was begun in the summer of 1937 as part of a project under a grant from the Penrose Bequest of The Geological Society of America to Dr. M. H. Stow Financial assistance from the Yellowstone-Bighorn

Research Association made possible field work in July 1938. The field work was completed during the summer of 1939 with the aid of a grant from the Penrose Bequest of The Geological Society of America for which the writer is duly grateful. Thanks are due Professor W. T. Thom, Jr., who suggested the problem, and Dr. Erling Dorf, who identified several plant fossils. The writer is especially indebted to Dr. Marcellus H. Stow for numerous profitable discussions and suggestions concerning the field work. George Phair was a capable field assistant in 1939 and did part of the petrographic work on the Iron Mountain intrusive as an undergraduate thesis at Hamilton College. The writer also wishes to thank Henry and Jack Eagle, Joe Kerr, Robert Andrews, and John Weitz for their very able field assistance.

LIVINGSTON FORMATION

The name Livingston formation was first applied by Weed (1893) to about 7000 feet of sandstones, shales, and conglomerates north and east of Livingston, Montana. Later studies by Stone and Calvert (1910) and others have shown that this formation grades laterally into the Claggett, Judith River, Bearpaw, Lennep, Lance, and Fort Union formations and has resulted in considerable remapping of Livingston formation areas. Present-day use of the term Livingston formation (Vhay, 1939) often refers only to Weed's (1893, p. 26-29) agglomerate member of the Livingston,—a member mapped as basic andesitic breccia in the Livingston Folio (Iddings and Weed, 1894) and including all the pyroclastic rocks discussed in this paper. Considering these changes in the use of the name Livingston, the writer has called these pyroclastics the Livingston igneous series to avoid further confusion.

PYROCLASTIC ROCKS

TYPES AND EXTENT

The Livingston igneous series comprises breccias, tuff-breccias, lapilli-tuffs, tuffs, vent agglomerates, tuffaceous sandstones, and volcanic conglomerates. Throughout this paper the writer uses the classification and terminology of pyroclastic rocks suggested by Wentworth and Williams (1932, p. 45-51). The areal extent of this series is shown on Figure 1. This pyroclastic outcrop corresponds approximately to the agglomerate area on the map of Stone and Calvert (1910, Pl. 7, facing p. 555).

Breccias, occasional tuffs, and vent agglomerates make up most of the series in the area at the headwaters of Upper and Lower Deer creeks and Bridger Creek (Fig. 2). These breccias show essentially no bedding. The fragments vary greatly in different beds but average 2 to 3 inches in diameter with a maximum of 3 feet locally. The matrix of the breccias is usually a crystal tuff. Tuffs and lapilli-tuffs are not numerous. These pyroclastics are cut by breccia-filled plugs, and vent agglomerates.

Eastward along the Stillwater River near Nye (Fig. 1) the series contains well-bedded breccias, tuffs, and intercalated tuffaceous sandstones (Vhay, 1939, p. 434). Northward toward the Yellowstone River and westward along the Boulder River well-bedded volcanic conglomerates and tuffaceous sandstones predominate. Due to their lenticular character single beds do not extend far. Cross-bedding is not uncommon in the tuffaceous sandstones.

This pyroclastic series is at least 2000 feet thick in the Deer Creek headwaters region and thins to the north and west to approximately 50 feet at the Yellowstone River.

COMPOSITION

The breccias, tuffs, and volcanic conglomerates are all andesitic. Coarse and fine fragments alike consist of augite andesite, hornblende andesite and less commonly dacite and basalt. The andesites are composed of augite, hornblende, labradorite, andesine, minor biotite, magnetite, apatite, and orthoclase, with augite, hornblende, and labradorite as phenocrysts. The matrix is composed of fragments of these same minerals. Most of the pyroclastics contain fragments of all these types suggesting admixing from various contemporary vents.

The Enos Mountain breccias are exceptional, for they consist almost entirely of biotite dacite and have occasional limestone fragments. The dacite is composed of biotite, hornblende, andesine, orthoclase, quartz, magnetite, and apatite, the first three occurring as phenocrysts.

A few thin basalt and andesite flows are locally interbedded with breccias near vents. Undoubtedly more exist than were actually found, but even so, flows probably make up less than 1 per cent of the extrusive rocks.

AGE

Leaves and stem fragments are found sparingly in some of the tuffs and volcanic sandstones, and irregular pieces of silicified wood occur rarely in the breccias. Dr. Erling Dorf (written communication) has identified the following species of leaves from a tuff bed at the base of the volcanic series on the west branch of Upper Deer Creek:

Pteris nitida Hollick. One good specimen similar in observable details to specimens of this species from the upper? member of the Chignik formation of Alaska (Montanan age, Upper Cretaceous).

Sequoia reichenbachi (Geinitz) Heer. One excellent specimen indistinguishable from the specimens of this species from the Upper Cretaceous Judith River formation

Dryophyllum subfalcatum Lesquereux. Two good specimens. Widespread in post-Coloradan Upper Cretaceous.

Rulac quercifolium Hollick. Five good specimens referable to this species from the middle and upper members of the Chignik formation of Alaska (Montanan age, Upper Cretaceous).

Viburnum montanum Knowlton. Five excellent specimens. Widespread in post-Coloradan Upper Cretaceous.

Vitis stantoni (Knowlton) Brown. Three excellent specimens similar in all respects to the type specimen of this species (called *Castalia stantoni* Knowlton) from the Judith River formation. These are quite different from reported occurrences of this species in the higher Upper Cretaceous.

Carpolithus cf. *C. carrollensis* Berry. One excellent specimen very close to this Upper Cretaceous species from the Ripley formation.

? *Marchantites,* new species.

Dorf (written communication) says: "This combination of species is quite clearly an Upper Cretaceous assemblage of Montanan age. I believe it to be approximately equivalent in age to the Judith River formation of central Montana."

ORIGIN OF EXTRUSIVE SERIES

GENERAL STATEMENT

The Livingston igneous series originated in the intermixing of ejectamenta from a large number of small contemporary vents. During and after the eruptions, the pyroclastic material was modified first by mud flows and later by streams.

VOLCANIC VENTS

Several small vents 1000 to 3000 feet in diameter, filled with agglomerate or breccia occur in the area at the headwaters of the Deer creeks (Fig. 2). The vent agglomerate or breccia in half of these pipes is not much coarser than that in the surrounding areas. However, at least three of the vents were distinguished by fragments averaging 2 to 3 feet in diameter. In one vent blocks more than 6 feet across are numerous. All these vents show vertical jointing and are cut irregularly by intrusive pipes, indicating successive periods of intrusion. It is not possible to tell which part of the extrusive series came from a particular vent, except at the Enos Mountain volcano.

The Enos Mountain volcano (Rouse, *et al.*, 1937, p. 724), a compound dacite breccia plug, is the largest known vent. However, the pyroclastics originating from this vent are limited to 3 or 4 square miles of outcrop area and are only a few hundred feet thick.

EXPLOSION BRECCIAS

The volcanic breccias in the area of the vents, being poorly bedded and having angular fragments embedded in a crystal tuff matrix, appear to be explosion breccias of vulcanian type eruptions. However, the absence of good cone structures indicates post-eruption rearrangement of the ejecta-

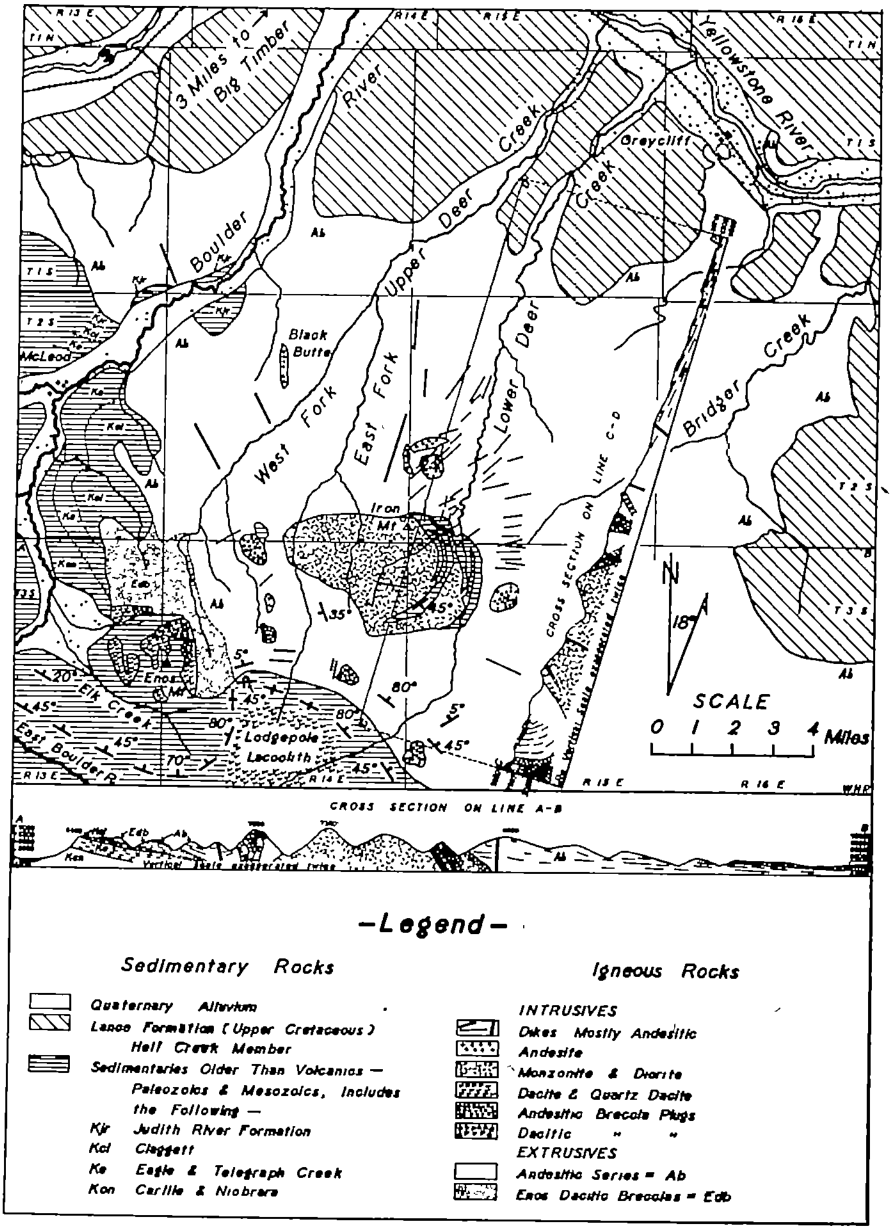

FIGURE 2.—*Geologic map and cross sections of the Deer Creek-McLeod area, Montana*

menta; dry avalanches or mud flows obscured part of the original bedding before consolidation.

MUD FLOWS AND STREAM DEPOSITS

The author believes that the breccias just north of the vents and along Bridger Creek and the Boulder River were deposited chiefly as mud flows because the fragments are heterogeneous and subangular. Vhay (1939, p. 436-437) has suggested that the breccias and tuff-breccias near Nye and just east of the Deer Creek area were deposited chiefly as hot and cold mud flows.

Still farther from the vents to the north and west where volcanic conglomerates and sandstone become numerous, the lenticular beds resemble stream deposits. Deposition by numerous, shifting, overloaded streams on an alluvial slope is suggested. However, even here the matrix of the volcanic sandstones is composed of angular crystal fragments.

INTRUSIVE ROCKS

GENERAL CONSIDERATIONS

The intrusive rocks of the Deer Creek area consist of a small laccolith, many dikes, and a few pipes in addition to the breccia plugs and vents. The Lodgepole laccolith (Rouse, *et al.*, 1937, p. 721-724) belongs to this group (Fig. 2).

IRON MOUNTAIN LACCOLITH

General statement.—An intrusive body 3 to 4 miles across, Gold Hill of Rouse, *et al.* (1937, p. 726), includes Iron Mountain (Fig. 2). This body has raised Paleozoic and Mesozoic sediments, Madison limestone to Cloverly conglomerate inclusive, around its north and east sides to dips of 45° (Fig. 2). A few marginal sills of varying thickness give this body the general form of a Christmas-tree laccolith similar to the Lodgepole laccolith (Rouse, *et al.*, 1937, p. 721-724).

The laccolith is a fine-grained granitic rock varying from augite diorite to hornblende monzonite. The former consists essentially of andesine, augite, magnetite and biotite with accessory apatite, while the latter contains green hornblende, albite-oligoclase, and orthoclase with minor biotite, quartz, magnetite, apatite, and sphene. The biotite is slightly altered to chlorite, and the orthoclase to sericite or kaolin.

Contact metamorphism.—A few small unimportant lenses and masses of magnetite and skarn are located at or near the contact of the monzonite and Madison limestone. Such lenses parallel the bedding of the limestone and reach a maximum length of 20 feet and width of 2 feet. This local contact metamorphism shows chiefly fine-grained magnetite (some lodestone), yellowish-brown massive garnet, and fine-grained epidote with minor specularite, wollastonite, and quartz.

Hydrothermal veins.—A number of narrow hydrothermal fissure veins and joint fillings near the center of the intrusive mass have excited prospectors for half a century. These veins, never over 2 to 3 feet wide and a few hundred feet long, consist essentially of calcite and a fine-grained mixture of sericite and quartz with a little pyrite. The latter is reputed to carry traces of gold.

Near these veins the monzonite and crosscutting porphyry dikes show more alteration to chlorite, sericite, and kaolin than elsewhere in the laccolith. This hydrothermal alteration suggests the location of the pipe which fed the laccolith.

DIKES AND PIPES

About 100 narrow vertical dikes radiate from the laccolith. The majority of these are 4 to 10 feet wide, often traceable for many hundreds of feet, and are localized to the north and east of the laccolith (Fig. 2). Northwest of this center, the few dikes present are much wider; the one holding up Black Butte is well over 100 feet wide. A few dikes actually cut the laccolith.

One thick, faulted, semicircular dike north of Iron Mountain dips inward about 45° toward a breccia-filled plug after the manner of a cone sheet.

These dikes are all andesite porphyries with phenocrysts of plagioclase (labradorite and andesine), hornblende, augite, magnetite, and minute apatite crystals in an aphanitic groundmass. About one third are hornblende andesites, one third augite andesites; in the other third hornblende and augite occur in equal amounts. A little biotite is present in about one third of the dikes.

A few small irregular andesitic pipes occur in the breccia plugs as part of the vent structures. An isolated vertical pipe about 1000 feet in diameter is a latite containing about equal amounts of orthoclase and plagioclase with very minor biotite, augite, and magnetite.

STRUCTURE

RELATION TO OLDER ROCKS

The Livingston igneous series overlies with slight angular unconformity the Eagle, Claggett, and Judith River formations as exposed along the Main Boulder and East Boulder River valleys. This unconformity represents a surface of low relief and dips gently northward from the vicinity of Enos Mountain reaching stream level a few miles north of McLeod in the Boulder valley.

On the east side, in the Stillwater area (Fig. 1), the same angular unconformity was mapped by Wilson (1936, p. 1170), Vhay (1939, p. 434), and Stow (written communication).

Several intrusives (Vhay, 1939, p. 434) outcrop south of the volcanic series, principally the Lodgepole laccolith (Foote *in* Rouse, *et al.*, 1937, p. 721-723). These bodies have tilted Paleozoic and Mesozoic sediments to near vertical. Later the main Beartooth thrusting has further deformed these beds. Some of the pyroclastics have also been tilted particularly near the headwaters of Lower Deer Creek. Here the volcanics are nearly conformable with Upper Cretaceous beds of Coloradan or Montanan age. However, some of the extrusive beds overlie unconformably the upturned sediments particularly near the headwaters of Upper Deer Creek where nearly horizontal breccias overlie vertical and overturned Cretaceous beds, Cloverly and younger. Thus some breccias postdate the intrusion of the Lodgepole laccolith, although others are older and were tilted by the doming of this laccolith. The Lodgepole laccolith was intruded prior to Beartooth thrusting (Rouse, *et al.*, 1937, p. 722).

RELATION OF YOUNGER ROCKS

Northward and northeastward the gently north dipping volcanic series is overlain conformably by Lance sediments. The bed immediately overlying the volcanics from the Boulder River area near Big Timber around to the Stillwater River area has been identified as the Hell Creek member of the Lance by Stow (written communication).

STRUCTURES WITHIN THE VOLCANIC SERIES

The lack of good bedding in the volcanic series of the vent area, due to dry avalanches and mud flows, has made it impossible to identify any actual cone structures around the various vents and difficult to distinguish dips of deposition from deformational structures. Fortunately, sufficient bedding was present to indicate the relationships at the south edge of the Deer Creek area.

The relation of the breccias to the Iron Mountain laccolith is rather obscure. Certainly much of the breccia is later than this laccolith and overlies it and the upturned sediments unconformably. Some steeply dipping but poorly bedded breccias near the southwest side of the laccolith suggest that some pyroclastic material was domed by the emplacement of the Iron Mountain laccolith.

Two angular unconformities have been recognized within the volcanic series. The Enos Mountain dacitic breccias are overlain by andesitic breccias with a slight angular unconformity. At least one similar unconformity exists within the andesitic beds, best seen north of Iron Mountain and east of Lower Deer Creek where truncated breccias dipping 45°

NE. are overlain by nearly horizontal breccias; this steep dip may be partly depositional. Periods of eruption seem to have been separated by periods of considerable erosion and some tilting and deformation.

SUMMARY OF CONCLUSIONS

GENERAL CONSIDERATIONS

Geographically and genetically the Livingston pyroclastic and intrusive rocks are a distinct unit of the western Montana upper Cretaceous volcanic field.

The pyroclastic material was exploded from a large number of small vents and was modified by mud flows and streams.

Intrusive rock types include andesite, latite, dacite, diorite, and monzonite. Augite and hornblende are almost equally abundant, and biotite in small amounts is characteristic. Quartz is absent or rare. The pyroclastics contain fragments of these same rocks and minerals, though andesite greatly predominates.

SUMMARY OF EVENTS

(1) The first extrusive activity in the Deer Creek area occurred in Judith River time as indicated by leaf fossils. The Enos Mountain dacitic vent was one of these early extrusive centers. As eruptions occurred in some areas, Judith River sediments were being deposited simultaneously nearby.

(2) Slight erosion of dacitic pyroclastics and Judith River sediments

(3) Widespread andesitic breccias were deposited unconformably on dacitic breccias and Judith River sediments. Periods of volcanic activity alternated with periods of erosion and redeposition.

(4) Intrusion of the Lodgepole and Iron Mountain laccoliths domed Paleozoic and Mesozoic sediments as well as the earlier breccias.

(5) Widespread erosion of the earlier vents and pyroclastics uncovered the Iron Mountain intrusive rock. The products of this erosion were deposited to the north and west as tuffaceous sandstones and volcanic conglomerates.

(6) Extrusive activity occurred from a large number of small vents. This later material was deposited unconformably on the tilted and beveled edges of the earlier pyroclastics.

(7) Intrusion of the majority of the andesitic dikes most of which cut the later breccias.

(8) Volcanic activity had ceased by Lance time. The Hell Creek member of the Lance overlies conformably the volcanic conglomerates, tuffaceous sandstones, and breccias to the north and east.

(9) The main Beartooth thrusting followed this volcanic activity but did not affect the volcanic series appreciably.

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TIME RELATIONS IN OCEAN SEDIMENTS

BY C. S. PIGGOT AND WM. D. URRY

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ABSTRACT

Nonequilibrium between the radioactive elements in the surface of the ocean bottom provides a method of measuring time in the past 300,000 years. The principles of this method are not to be confused with those underlying the radioactive methods of determining geologic time measured in millions of years The well-preserved geological and biological history of the ocean basins has been worked out for many of the cores of ocean sediments procured by Piggot, and the above method of measuring time intervals in years has been applied to these cores. The results indicate: (1) that the effects of glaciation on the continents are contemporaneous with equivalent effects on the type of deposit in the ocean bottom; (2) that the effects of glaciation on the type of ocean sediment are widespread, extending in the northern hemisphere at least to the Caribbean Sea; (3) that within short intervals of time there is, in the ocean bottom, a considerable variation in the type of deposit attributable directly or indirectly to climatic changes which, probably because of the continuous effacement of the evidence on land, has not been reported in studies of continental glaciation.

This method cannot compete in accuracy with the use of varved clays, but it possesses the advantages of (1) application to a considerable area of the earth's surface, (2) reference to the present time, and (3) applicability to at least half of the Pleistocene epoch

The rate of deposition of ocean sediments is treated as a subsidiary problem. The obvious step from the determination of time intervals in a core to the rate of deposition of the sediment represented by the core is complicated by the present incomplete knowledge of the distortion of the sediment in the process of taking the sample. The results in general agree with previous estimates by other methods, some of which, however, are open to the same criticism It is possible to discuss within limits the variations of the rate of deposition of ocean sediments in the past.

FOREWORD

The lithology, biology, and chemistry of cores of ocean-bottom sediments have been described in the literature. The comprehensive examination by Bradley and associates (Bradley *et al.*, 1940) of the geology, chemistry, and biology of a suite of North Atlantic deep-water cores obtained by Piggot in 1936 constitutes a great advance in such researches, primarily because the cores were considerably longer than those previously available, and because of the wealth of detail that has been accumulated by a number of specialists working on identical material.

The concentrations of the radioelements in these cores have been determined, and the radioactive relations thus established offer a method of dating any horizon in these vertical sections of the ocean bottom. The geological and biological studies indicate several important datum planes, which provide added interest in an objective method by which core lengths may be converted into time intervals.

Measurements of the time intervals have been made in some cores not included in the North Atlantic group. The Foraminifera from one of these cores from the Bartlett Deep in the Cayman Trough have been studied by Cushman (1941), and his findings stimulated the present work on this core.

In order that a given date may have significance, some event must exist; the reports, to which reference is made above, provide many such events.

ACKNOWLEDGMENTS

Thanks are extended to Dr. Joseph A. Cushman for helpful discussion and communication of unpublished results of foraminiferal studies in these cores; to Dr. Roger Revelle of the Scripps Institution of Oceanography, La Jolla, California, for supplying many specimens of a core of Red clay; and to Dr. Juul Hvorslev for helpful criticism and discussion of the problem of distortion in the process of taking samples of the ocean bottom.

INTRODUCTION

Heretofore there has been no direct method of dating the many events that have left their records undisturbed in the ocean bottom.

Rate of deposition of ocean sediments has been estimated by assuming a date for an observed event, as, for example, the estimate that the accumulation of postglacial warm-water ooze began in the equatorial Atlantic 20,000 years ago. By this method Schott (1935, 1939) calculated the average rates of accumulation of the three main types of ocean sediments. Lohmann (1909) based similar estimates on the population and rate of growth of the Coccolithophoridae, and Twenhofel (1929) estimated the amount of inorganic material deposited on the sea floor remote from the coasts. The estimates by Schott and the measurements reported here indicate that the fluctuations over any considerable area are so great as to invalidate the timing of events at one place in the ocean by the data obtained from another locality.

It is now known that the core length does not bear a one to one, or uniform, ratio to the depth of the sediment sampled (Piggot, 1941, Emery and Dietz, 1939; 1941), and determinations of rates of deposition of the sediments are accompanied by uncertainties that are difficult to evaluate. This applies to such estimates as have been made previously and also to the values given here. These uncertainties do not exist in the present method of dating a particular event, the evidence for which is found at a given core depth, and to the authors this seems to be the more immediate problem.

Any direct observations of the rate of deposition of sediments as made by Murray and Peake (1904) are subject to considerable experimental error and pertain only to the present time and place of observation. Lack of knowledge of the rate of deposition of all classes of sediments during the past seriously hinders the use of the geological sediment column as a time scale (Schuchert, 1931). For this reason, the physical properties of the radioelements, whose rates of disintegration are invariant, have been used to estimate geologic time. Since helium and lead are the ultimate stable products of the radioactive disintegrations of uranium and thorium, the former elements have become associated with two established methods of determining geologic time (Lane, 1924-1941). While the principles of these methods are simple, the applications are subject to inherent uncertainties.

For measuring recent intervals of time in the Pleistocene, some properties of the radioelements can be employed which do not depend upon the accumulation of the stable end products. This results in a method that is largely free from the uncertainties of the "helium" and "lead" techniques, which in any case are not applicable to the immediate past.

The method has been aptly termed by Lane (1937) the "Per cent of Equilibrium Method," but its principle has been recognized since the early days of the science of radioactivity. It has received scant attention because it is limited to the past half million years.

METHOD

Radium is the third element of importance in the series of disintegrations by which uranium is transformed into lead. If a geological specimen has remained chemically undisturbed throughout the Pleistocene or longer, the elements in this series will be in "radioactive equilibrium," and the relative amounts of each will be fixed and independent of time. If, in the course of formation of a more recent specimen, radium should be segregated from its progenitors, its concentration in such a specimen will decrease exponentially with a half-life of 1700 years. This simple example of the method will rarely apply in nature, although Schlundt (1934) apparently found it in the tufas of Yellowstone Park.

A more complicated case exists if the uranium is completely segregated from its daughter and granddaughter elements ionium and radium. For practical purposes, the uranium may be considered as an inexhaustible reservoir. It cannot produce radium directly; it must first disintegrate into ionium, which has a half-life of 82,000 years. Consequently, the radium content will increase with time but at a considerably slower rate than it decreases in the simple case treated above. This provides a more promising means of measuring time in the Pleistocene than that depending upon the disintegration of radium by itself, because after 300,000 years there is still a detectable difference between the radium concentration and the ultimate constant value that it will attain when radioactive equilibrium has been established. The theory of this case has been treated by Kovarik (1931). Such clean-cut separations seldom occur in nature, and their existence would be difficult to substantiate.

Cores of ocean-bottom sediments, which rarely are disturbed by tectonic or other physical forces, provide a means of following the concentrations of the radioelements as these change with time. Such studies (Piggot and Urry, 1939; 1941; 1942; Urry and Piggot, 1942) have revealed that all three of the radioelements mentioned above are being incorporated in the structure of the ocean sediments at any certain place. The ratios of the concentrations of uranium, ionium, and radium being deposited differ greatly from the fixed ratios in radioactive equilibrium, and thus there is a complicated variation of the radium content with depth below the surface of the sediment. The ocean sediments as laid down contain very little uranium, considerably more ionium than is equivalent to this uranium, and, in most cases, less radium than that appropriate to equilibrium with the ionium.

In Figure 1, curve A illustrates the change of the radium content with time. The ionium first produces radium at a rate in excess of that at which the radium decays, therefore the radium content increases until these rates are equal. The time required to accomplish this is established by the laws of radioactive disintegration, and it is dependent upon the ratios of the initial amounts of radium, ionium, and uranium. The whole of curve A is equally dependent upon these ratios, and it must therefore be calculated for each core. After the attainment of equilibrium with its parent ionium, the radium concentration is controlled by the decay of the former element. The ionium concentration decreases because of insufficient uranium to support it. Finally, the number of atoms of ionium disintegrating will become just equal to the number

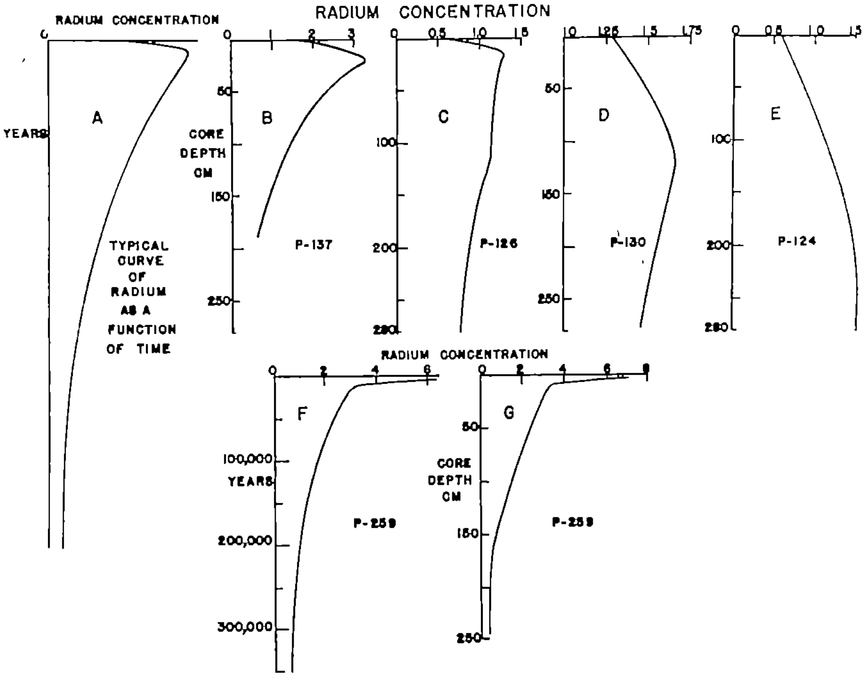

Figure 1—*Variation of the radium concentration below the surface of the ocean bottom*

Curves A and F express the radium concentration as a function of time, calculated from the equations of radioactivity for two sets of initial conditions Curve F is calculated for the actual initial conditions found experimentally (curve G) in core P-259. Curve A is a typical curve which must be calculated for each of the specific cases represented by the experimental curves B, C, D, and E The radium concentration is expressed in 10^{-13} g. radium per cc of the original sediment—i.e., it is the radium content in a sample of standard cross section (1 sq cm) and unit depth (1 cm) of new ionium atoms produced from the small amount of uranium present, and radioactive equilibrium among all three elements will be established. This is illustrated in Figure 1, curve A, which shows the attainment of an essentially constant radium content after the lapse of 300,000 to 400,000 years. Curve F in Figure 1 is the theoretical curve of radium as a function of time for a core of Red clay, and curve G is the experimental curve of radium content as a function of depth in the same core. This Red clay differs from the other sediments examined in that radium was deposited in excess of that which is appropriate to equilibrium with the ionium deposited. Why this should be so is not yet known, but chemically and biologically the Red clay differs greatly from the calcareous blue muds and Globigerina oozes which comprise the other examples in Figure 1.

The analysis of each curve in order to convert units of core depth into time intervals necessitates superposing the appropriate curve A on the experimental curve. The mathematical treatment of such complex mixtures of the radioelements and the method of accomplishing this conversion have been given by Urry (1942). It is not possible to superpose the curves graphically, and the somewhat lengthy mathematical process cannot be discussed here. However, the procedure may be studied qualitatively from Figure 1. Curve B needs little explanation because its profile is so similar to that of the theoretical curve A. In curve C, the more gradual decrease in the radium content indicates that a much shorter period of time was required to deposit 280 cm. of core P-126 in the North Atlantic than to lay down the 190 cm. of core P-137 in the Cayman Trough of the Caribbean Sea. Still shorter periods of time are involved in cores P-130 and P-124 where the maximum radium content is about half way down, and near the bottom of each core, respectively. Curves B, C, D, and E therefore represent decreasing intervals of time required for the deposition of the sediments represented by the respective cores. Changes in the rate of deposition are shown by a departure of the experimental curve from the appropriate theoretical curve. Failure of the radium to decrease normally for a certain portion of the core indicates an accelerated rate of deposition for a certain time, whereas an abnormal decrease in the radium content indicates a retarded rate of sedimentation. For instance, curve C, for P-126, indicates an abnormality in the rate of deposition between 80 and 120 cm. from the top of the core, which will be discussed in the next section.

The experimental curve G for the Red-clay core, P-259, demonstrates the attainment of complete radioactive equilibrium as shown by the constancy of the radium content below about 160 cm. In this region the radium content provides a measure of the uranium content which for all practical purposes is constant over the interval of time represented by any of the cores. The initial ionium content can be determined from the slope of that portion of the curve which represents the decay of the ionium. Thus it is possible to acquire a complete knowledge of the relative amounts of uranium, ionium, and radium at any time merely by measuring the relative concentrations of radium at a series of depths in a core. It is not even necessary to refer the determinations to a radium standard. This has been done, however, because the radium determinations are of use for other purposes besides the dating of events in the sediments.

The cores represented by curves B, C, D, and E, although as long as P-259, are far from representing the interval of time required to establish radioactive equilibrium. In such cases the uranium must be measured by an independent method. Fortunately, a knowledge of the uranium content is imperative only when the core represents a considerable fraction of the time required to attain equilibrium, as is the case with P-137. An independent measurement of the uranium content of this core has been made by Urry (1941). The uranium content is quite unimportant in the analyses of curves D and E, in Figure 1.

It must be emphasized that the conversion of the core depth in centimeters, into time in the past—*i.e.*, the superposing of the theoretical curve A on the experimental curve B—requires the assumption that the sediment has acquired a constant concentration of the radioelements from the sea water or other sources throughout the period of its deposition. This need only be true of the deposit at the location of the core. Quite appreciable short-period fluctuations will cause but little error provided these smooth out within 100 years or more of sedimentation. The following facts support this assumption. In cores of uniform composition the experimental curves are quite smooth. The uranium content of core P-137 is constant throughout the length of the core, within the limits of error of the measurements (Urry, 1941). If the variations of the radium content, as shown in the experimental curves, are due to unknown causes quite unrelated to the established laws of radioactivity which give curves A and F, then the unknown causes must produce these variations according to equations analogous to those governing radioactive decay. The authors have examined this possibility but find no basis for such a contention.

Core P-126 contains several zones of glacial marine deposits where the character of the sediment is different from the warm-water zones (Bradley *et al.*, 1940). As might be expected, the radium concentrations in these coarser sediments depart, although not seriously, from the curve drawn through the points for the normal sediment of the core. For this reason, the analysis of the cores from the North Atlantic must be more detailed than those of the sediments from lower latitudes (P-137, P-259), which are known to be more uniform in character.

TIME INTERVALS IN OCEAN-BOTTOM CORES

NORTH ATLANTIC CORES

The complete suite of ten cores obtained by Piggot on board the cable ship LORD KELVIN from the North Atlantic has been examined lithologically and geologically by Bramlette and Bradley (Bradley *et al.*, 1940, Pt. 1), and the Foraminifera have been studied in detail by Cushman and Henbest (Bradley *et al.*, 1940, Pt. 2). The three cores from this suite chosen for the preliminary radioactive investigations showed the greatest difference in the relative depths of the zones as established by the lithological and foraminiferal studies. The disposition of the zones has been plotted in Figure 2 against the core depth. The key to the shading is that employed by Bramlette and Bradley (Bradley *et al.*, 1940, Pl. 3), with the exception of the zones of anomalous deposits. The numbers opposite the lines defining the zones or the temperature dots give the intervals of time in years measured from the top of the core as zero.

While it is unnecessary to redescribe the features of each zone, which can now be assigned a date instead of a depth in the core, attention may be drawn to a few important points.

That the time interval represented by the North Atlantic cores varied considerably was obvious even before the results described here were available. This was demonstrated not so much by the presence and number of glacial zones [which need not be contemporaneous, if Bramlette and Bradley's (1940, p. 7-9, Fig. 2) interpretation of their formation be accepted] as by the position of the well-defined base of the zone of alkalic volcanic glass shards which is part of the uppermost warm-water zone in all the cores except No. 11. Further evidence was provided by a lower volcanic zone in all the cores [except P-124(3)] west of the Mid-Atlantic Ridge. This zone was not penetrated even at greater core depths east of the Ridge. The second volcanic zone is shown in core P-126(5) in Figure 2. In core P-124(3) no glacial zone was penetrated, and the base of the upper zone of volcanic shards was not reached

These shards are assumed to have originated in explosive volcanic eruptions and to have been transported initially through the air. They are more numerous at the bottom of the layer, which is sharply defined so that their deposition implies contemporaneity over large geographical areas. This is confirmed by the occurrence of the base of the upper ash

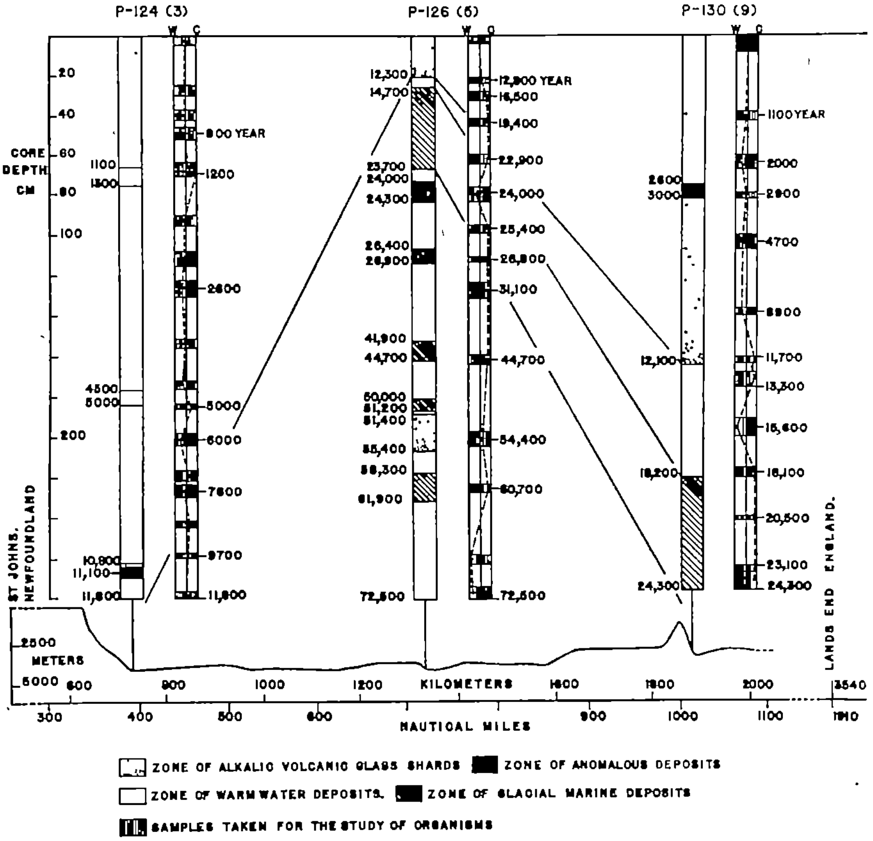

FIGURE 2.—*Chronology of sediments in the North Atlantic*

The dates of various events (present time = 0) which have left evidence of their occurrence in the ocean sediments of the North Atlantic. The vertical scale is in centimeters of core depth· the horizontal scale in distances. Each core is represented by two profiles. The left-hand profile shows the lithological events (Bradley *et al.*, 1940, Pl. 3), and the right-hand profile shows the past fluctuations in the temperature of the surface water at the site of each core (Bradley *et al.*, 1940, Pl. 4). The dot in each shaded block indicates qualitatively the deviations from the present temperature. The station numbers in parenthesis after each laboratory core number are the same as those used by Bradley (1940).

The numbers against the lines defining the zones or opposite the temperature dots give the intervals of time in years measured from the top of the core as zero. The quotation of time intervals to the nearest 100 years must not be taken too literally. Such accuracy of statement is probably warranted in the case of P-124 (3) and for P-130 (9) above the zone of glacial marine deposits Some of the figures for core P-126 (5) are within this degree of accuracy, but little significance should be attached to differences of less than 1000 years below the second glacial zone in this core

zone at 12,000 years in both P-126(5) and P-130(9). The fact that core P-124(3) represents no more than 12,000 years places it entirely in the upper volcanic zone, as is indicated by the lithology. According to the dating of the second volcanic zone in core P-126(5), similar ash was deposited between 51,000 and 55,000 years ago and was followed by conditions which gave rise to 1000 years of glacial marine deposition.

The zones that are dated in core P-124 are three thin and rather widely spaced layers of less limy mud (Bradley *et al.*, 1940, p. 5). Cushman and Henbest (Bradley *et al.*, 1940, Pl. 4) find some indications of colder surface waters during the deposition of these clay zones. The anomalous zone of terrigenous silt at 11,000 years presumably was caused by slumping (Bradley *et al.*, 1940, p. 16). Between 2000 and 4500 years ago, and again at about 6000 years ago, in core P-124(3), there is some indication that the surface water was warmer than at present. Core P-130(9) indicates that a warmer period existed from 2000 to 9000 years ago, and also about 16,000 years ago.

Four zones of glacial marine deposits were penetrated in core P-126(5). This is the maximum number found in any of the cores. All four zones occur within the past 62,000 years. These zones in core P-126(5) have been plotted against time in Figure 3, and all are contemporaneous with the Würm glacial age. The chronology of the Pleistocene for the Alps given in Figure 3 is that of Penck and Brückner (1901-1909). Kay (1931) has made a detailed study of the chronology of the Pleistocene in Iowa. His results are reproduced in Table 1 because of their bearing on the problem in the North Atlantic and elsewhere. The glacial ages in Iowa are brief because they represent only the most southerly extremities of the ice sheets, and therefore the time when the respective glaciations were at a maximum.

Kay's values are plotted at the top of Figure 3. The earliest glacial marine zone in core P-126(5) was deposited between 58,000 and 62,000 years ago. These dates are in reasonable agreement with Kay's proposal of a separate Iowan age between 55,000 and 58,000 years ago. Bramlette and Bradley (Bradley *et al.*, 1940, p. 11) favor the interpretation that all four glacial marine zones are substages of the Wisconsin stage. They do not mention an Iowan glaciation, either as a separate stage, or as a substage of the Wisconsin, but they discuss the early and late substages of the Wisconsin as proposed by Chamberlin and Salisbury (1906). The lower zone of volcanic ash above the earliest glacial marine zone may offer some justification for setting off the Iowan glaciation from the Wisconsin substages. Moreover, it is only in this interglacial marine zone in core P-126(5) that the temperature conditions approach those of the present day. This is substantiated by the temperature charts for the other cores 4, 6, and 7 (Bradley *et al.*, 1940, Pl. 4).

Contemporaneity of the earliest glacial marine zone with Kay's Iowan age implies contemporaneity of the Peorian loess with the lower volcanic ash zone. Kay (1931, p. 443) writes, "This loess was apparently laid down shortly after the deposition of the Iowan drift and not long before the deposition of the early Wisconsin drift." The base of the ash zone coincides with Kay's close of the Iowan age at 55,000 years.

It is interesting to note a considerable length of core between the last and next to last glacial marine zone indicated as a warm-water deposit

TABLE 1—*Minimum duration of the Pleistocene period in Iowa*

(After G. F Kay)

Interglacial age	Years	Glacial age
	0	
(Post late Wisconsin)		
	25,000	
		Late Wisconsin
	28,000	
Peorian		
	55,000	
		Iowan
	58,000	
Sangamon		
	178,000	
		Illinoian
	187,000	
Yarmouth		
	487,000	
		Kansan
	495,000	
Aftonian		
	695,000	
		Nebraskan
	705,000	

in core P-126(5), particularly since estimates of the time required for the recession of the continental ice fronts such as that of Antevs (1928) would indicate that this section of the core should be contemporaneous with the peak of the last glaciation. The temperature chart of Cushman and Henbest (Bradley *et al.,* 1940, Pl. 4), reproduced here in Figure 2 for the cores under consideration, indicates a continuation of cold-water conditions in this interval, and Bramlette and Bradley (Bradley *et al.,* 1940, p. 8-9) find the texture and composition of this zone to be intermediate between glacial marine deposits and foraminiferal marl, from which they infer a repeated migration of the limit of the drift ice back and forth across the site of these cores. The time represented by this zone in core P-126(5), when judged by its depth, is misleading. This is true of many core samples because of the distortion of the sediment introduced by the sampling procedure, as will be discussed under "Rates of Deposition." In this case much of the apparently long duration of the intermediate zone is due to two zones of anomalous deposits represented by the Bradley samples Nos. B-53 and B-55 (Bradley *et al.,* 1940, Pl 3). The upper zone, B-53, is 10 centimeters deep, and the lower zone, B-55, is about 8 centimeters deep. The radium content of both these zones is about the same and is considerably less than that indicated by the curve C in Figure 1 for the respective positions in the core; in fact, the radium content is comparable with that of the material between the earliest zone of glacial marine deposit and the lower volcanic ash zone. The time intervals as measured for these anomalous zones are extremely small and, within the limits of error, might be zero. This accounts for the curious shape of curve C in Figure 1. The accuracy of the method is insufficient to justify the conclusion that these anomalous zones originate by slumping, and the inference that the source of the slumped material was the sediment lying now at about 220 cm. in the core is unwarranted. Nevertheless, there is strong evidence for such a mechanism for the origin of the upper anomalous zone and very suggestive evidence for a similar origin of the second anomalous zone (Bradley *et al.*, 1940, p. 15-16). Bramlette and Bradley also suggest that the sediments of the lower zones of this core are a possible source of the material on the basis of lithological and temperature similarities.

Core P-130(9) provides an opportunity for detailed study of the features presented in the top 70 centimeters of core P-126(5). The contemporaneity of the boundary marking the presence of volcanic shards has been considered already. It is disconcerting to find that the deposition of the last glacial marine deposit ceased 3500 years earlier in this region than in the neighborhood of core P-126(5), although Cushman and Henbest find evidence for a brief cold-water period between 12,000 and 13,000 years ago, as shown in Figure 2. It is possible that the dates given for the lower portion of core P-130(9) are in error because the radium values for material in the glacial zone had to be given some weight. As previously mentioned, such values are subject to the influences of the source of the detrital material. In core P-126(5) there is evidence of four distinct periods of glacial marine deposition in 47,000 years, which requires considerable movement of the southern limit of the drift ice within a single glacial age, or two concurrent ages if Kay's Iowan be accepted. In the zone of intermediate character in core P-126(5), between 24,000 and 42,000 years, these northerly and southerly movements must have occurred in fairly rapid succession. Such a picture has been illustrated for a meridional section by Bramlette and Bradley (Bradley *et al.*, 1940, Fig. 2). The southerly limit of the drift ice, however, possibly varied at different longitudes at different times, so that the glacial marine deposits in the North Atlantic may not be completely contemporaneous.

GLOBIGERINA OOZE FROM THE CARIBBEAN SEA

Three cores from the bottom of the Cayman Trough in the vicinity of the Bartlett Deep were obtained by Piggot on board the ATLANTIS in 1937. The sediment is Globigerina ooze. The Foraminifera have been studied by Cushman (1941). Many interesting inferences have been drawn from these studies which make it more desirable to plot Cushman's relative abundance curves for the Foraminifera, as a function of time instead of core depth. It is first important to ascertain the correlation of these curves, which are believed to indicate swings of warmer and colder surface temperatures. The curves for the three cores P-135, P-136, and P-137 (Cushman, 1941, Figs. 5-10) show no correlation when plotted as a function of core depth. The method of dating cores described here offers an explanation. Each core was deposited in a very different interval of time. The radioactive measurements (Piggot and Urry, 1942) have not been analyzed completely, and the final results will be reported later. The results for P-137 are of interest. Cushman (1941) draws attention to the fact that ". . . the swings shown in core P-136 (Text Figs. 6, 7, 8), in particular, may be coincident with known glacial and interglacial periods." Cushman chose P-136 as probably representing the longest time interval on the basis of Schott's mean rates of sedimentation in the ocean and because of the well-marked periodicity of the abundance curves. The radioactive measurements indicate that the sediment represented by core P-137 required 300,000 years for its deposition, which is more than twice the time required for the deposition of the sediment represented by core P-136, although it is of similar length.

In Figure 3, some of Cushman's abundance curves are plotted as a function of time. The relative abundance of *Globorotalia menardii* has been studied by Schott (1933) as an index of surface-water temperatures comparable to those now prevailing in the equatorial Atlantic and therefore has been included. The duration of the Pleistocene after Penck and Brückner (1901-1909), and the duration of the Pleistocene ages in Iowa after Kay as given in Table 1, are plotted on the same time scale. The North Atlantic core P-126(5) is plotted on the same scale for the purpose already mentioned. There appears to be some correlation between the cold- and warm-water swings, as exhibited by the foraminiferal remains in core P-137, and the Pleistocene ages of glaciation. A more detailed examination of the Foraminifera in this core, between the positions corresponding to 55,000 and 100,000 years, was undertaken by Doctor Cushman at the writers' request, after the publication of his earlier results and after the time scale had been assigned. These additional data were required mainly to establish the correlations between cores P-136 and P-137, but it is interesting to note that cold-water zones, which were not evident from the first examination, but which were predictable from the present results for the North Atlantic cores and from Kay's results, have been found at about 55,000 and 65,000 years ago. However, the additional data indicate the existence of a brief period of colder surface water about 80,000 years ago, which has no apparent counterpart in the chronology of continental glaciation.

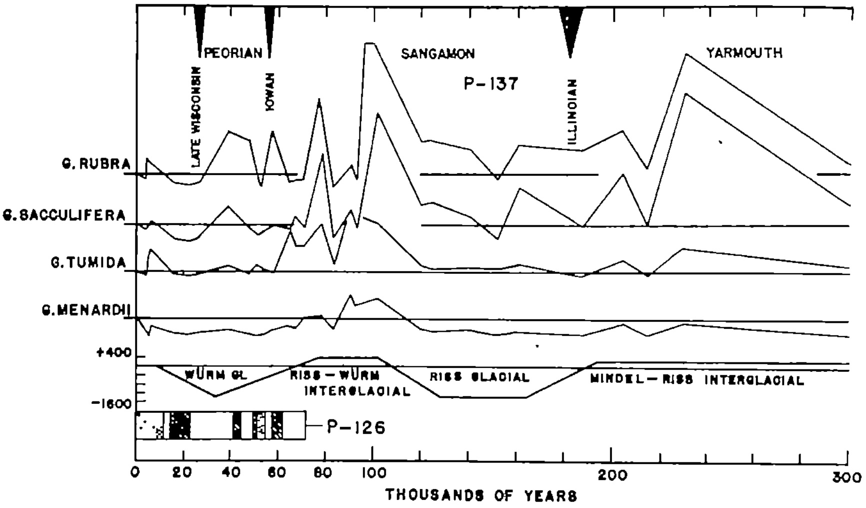

FIGURE 3.—*Changes in the abundance of Foraminifera with time in the Caribbean Sea*

The relative abundance of four of the pelagic species of Foraminifera studied by Cushman in core P-137 plotted against time. On the same abscissa of time are plotted the chronology of the Alpine glaciations after Penck and Bruckner, in terms of snow-line level in meters, the chronology of the Iowan glaciations after Kay (at the top of the diagram), and the chronology of the various zones in core P-126 (5) with the same key of shading as in Figure 2.

RED CLAY FROM THE PACIFIC OCEAN

Dr. Roger Revelle supplied specimens from a core of brownish red clay. The radium measurements on this material (Geophysical Laboratory No. P-259) were useful in illustrating the method earlier in this paper. The core was 246 cm. long, the upper 160 cm. of which represent 300,000 years. Below this, the method of dating becomes inaccurate because, as already explained, the radium concentration has practically reached a constant value in radioactive equilibrium with the uranium and ionium. If the rate of sedimentation at the depth of 140-160 cm. be applied to the sediment in the bottom of the core, the whole core represents a thickness of sediment requiring from 650,000 to 800,000 years for deposition. The wide limits are necessary because the distortion of the core during the process of sampling is difficult to appraise numerically.

RATE OF DEPOSITION OF OCEAN SEDIMENTS

GENERAL STATEMENT

In the previous section the authors have distinguished between the core and the sediment in the ocean bottom that the core represents. The length of the core is not the depth of the sediment that has been penetrated (core penetration), nor is this core penetration always equal to the instrument penetration—*i. e.*, the depth of penetration of the coring instrument.

From test sampling conducted by Piggot (1941) in the varved-clay pits near Hartford, Connecticut, it has been possible to establish the relations of these various depths. The data obtained in sampling the varved clays must be applied with caution to the cores of ocean sediments. A comparison of the mechanical properties of the varved clays with those of the cores discussed here indicates that the distortion of the ocean-bottom cores is perhaps somewhat less than that of the varved clays. The calculation of the rates of deposition of ocean sediments from the time intervals in ocean-bottom cores requires a knowledge of the depth of sediment represented by a given length of core. For the upper third of a core of varved clay this ratio is close to unity, so that the uncertainties involved in applying the data for the varved clays to ocean sediments probably do not exceed the errors of the measurements of the time intervals. However, this ratio of undisturbed sediment to core progressively increases and at the bottom of the cores of varved clay attains a value of 4 to 5.

INFLUENCE OF DISTORTION

The rates of deposition are presented here for the limiting cases. In Figure 4 is shown the rate of deposition of ocean-bottom sediments in the North Atlantic, as revealed by three cores from locations several hundred miles apart, the rate of deposition of a Globigerina ooze in the Caribbean Sea, and of a Red clay in the Pacific Ocean. The solid lines give the rates if there is no distortion of the sediment in obtaining the core, and the dashed lines give the rates when the depth of sediment is calculated from the core length with the distortion factors found for the varved clays. The true rate of deposition cannot be less than that indicated by the solid lines and is probably nearer to, but perhaps somewhat less than, that given by the dashed lines. This uncertainty, caused by a lack of knowledge of the distortion during sampling, applies also to the cores used by Schott (1935) to determine his oft-quoted estimates of the mean rates of deposition of the various types of ocean sediments. Most of Schott's cores were taken with a 20 mm. liner, which would introduce greater distortions than would the 50 mm. liner used by the authors.

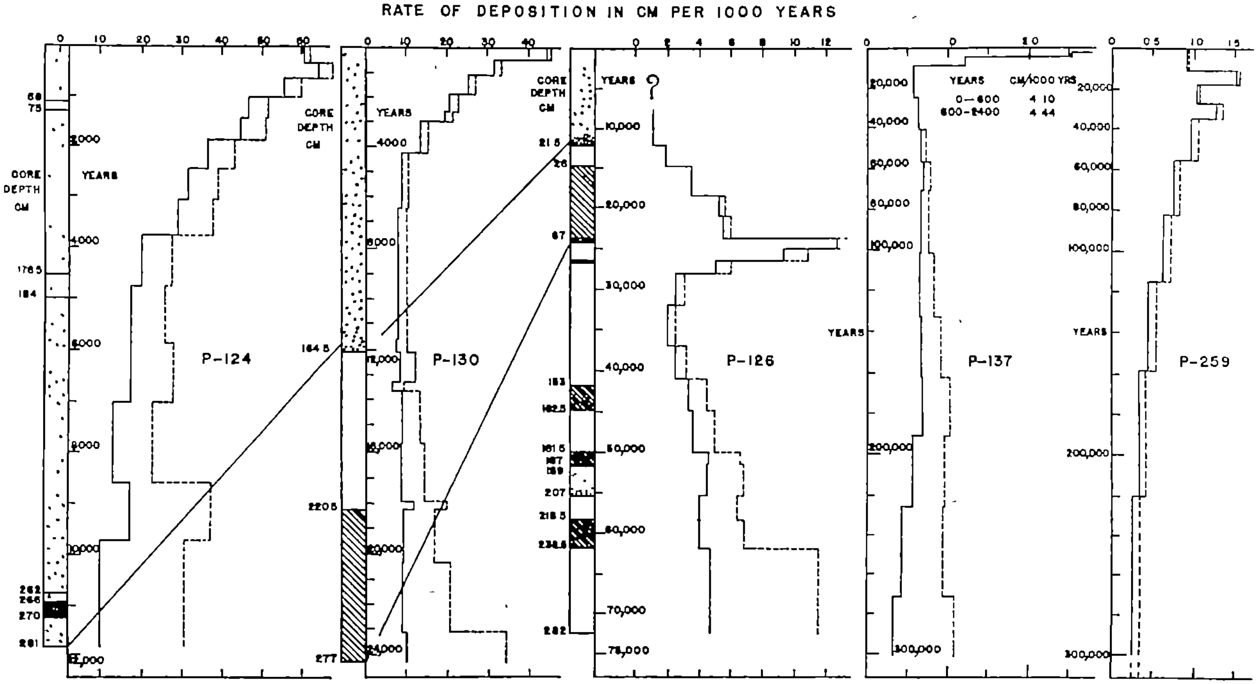

FIGURE 4—*Variations in the rate of deposition of ocean sediments in the past*

Some rates of deposition of ocean sediments in centimeters per 1000 years plotted against time. Solid lines are from values based on an assumption of no distortion of the sediment when sampling, dashed lines are from values based on the assumption that the distortion factors are the same as in the varved clays of Connecticut. The key to the profiles of the North Atlantic cores is the same as in Figure 2. Numbers to the left of each profile are core depths for important levels. The cross-reference lines for P-124, P-130, and P-126 are isochrones for 11,800 years, 12,000 years, and 24,000 years

Schott did not state that distortion had been considered. The diagrams of the cores given by Schott (1935, Beiläge I, II, III) show that the top of the layer which is free from *G. menardii* is close to the bottom of many of his cores. Recent measurements of the distortion in core samples indicate that the length of the core above this boundary must be considerably less than the depth of sediment sampled. Therefore, Schott's average values, expressed in cm. per 1000 years, are probably somewhat low. His values, however, are based on the assumption that the *G. menardii* began to accumulate 20,000 years ago in the postglacial time of the equatorial Atlantic. The uncertainty of this assumption is possibly as great as the error introduced by failure to correct for the distortion.

NORTH ATLANTIC SEDIMENTS

In Figure 4, the rate of deposition in cm. per 1000 years is plotted against time. The various zones in the three cores from the North Atlantic illustrated in Figure 2 are replotted against time instead of core depth, a few core depths being inserted for orientation. Isochrones connect points of correlated zones to illustrate the degree of contemporaneity. The time scales, from left to right, are in the ratios 1 : 2 : 6.25 : 25 : 25. Rates of deposition for the sediments represented by core P-124(3) are very high, either as based on the present measurements or on any previous estimates. This may be expected from the position of this core at the foot of the Grand Banks slope and in the Labrador Current. The rapid deposition of the sediments in core P-130(9) is similarly attributed to its relation to the Faraday Hills of the Mid-Atlantic Ridge.

Both cores give similar indications of the changes in the rate of sedimentation during the past 12,000 years. The rate of deposition in late postglacial time has increased and is unusually high at present. This is demonstrated in Figure 4 as applying to such widely separated areas as the Caribbean Sea, the Pacific Ocean, and the North Atlantic.

There seems to be some evidence (Fig. 4) that the deposition of the last glacial marine sediment was more rapid than that of the immediately ensuing warm-water deposit. On the other hand, deposition is more rapid after a prolonged period of warm surface water than during the periods of glacial marine deposit. The rates in the top of core P-126(5) are omitted in Figure 4 because, for any reasonable accuracy, the rapid change in the radium content with depth near the top requires the determination of the radium at closer intervals than is feasible. A high rate at present is indicated, however, as in the other North Atlantic cores. The lower part of the zone of intermediate character in core P-126(5), between the upper two zones of glacial marine deposit, is characterized by a low rate of sedimentation which is understandable, because the deposition of detrital material diminished and floor building was more dependent upon organisms existing at relatively low water temperatures. Similarly, following the last deposition of glacial marine sediments, there are indications that considerable time was necessary, after the prolonged period of cold surface water, to re-establish the normal abundance of organic life. This period may be measured in terms of some thousands of years. Most of the small fluctuations in Figure 4 are within the experimental error; in fact, the accuracy does not warrant tabulating the data in Figure 4.

The zones of anomalous deposits in core P-126(5) show abnormal rates of sedimentation. The rates as drawn in Figure 4 in the region of these zones were calculated after their thicknesses were subtracted, and they indicate that there was an increased rate of deposition, exclusive of the anomalous deposits, just prior to the last glacial marine sedimentation. The rates for the upper and lower anomalous zones, respectively, are 35 (39) and 15 (18) cm. per 1000 years. As previously mentioned, these high figures might well indicate almost instantaneous precipitation, because they are based on intervals of time so short as to be extremely dependent upon the drawing of the experimental curve. In this respect, the results support the theory of submarine slumping (Bradley *et al.*, 1940).

The absolute values for the rates of deposition in cores P-124(3) and P-130(9) must be considered as representative of abnormal conditions. Those for core P-126(5) probably indicate the order of magnitude for the general region west of the Mid-Atlantic Ridge.

CARIBBEAN SEA SEDIMENTS

The rate of deposition of the Globigerina ooze in the Cayman Trough (Caribbean Sea, core P-137) is roughly constant between 10,000 and 300,000 years at 0.5 to 1.0 cm. per 1000 years, but it increases during postglacial time to 4 cm. per 1000 years at present. The rates for this Globigerina ooze as given by the dashed curve are probably more accurate than those given by the solid curve, particularly for this core. If this is correct, the glacial ages are characterized by somewhat lower rates of deposition. This is contrary to the estimates of Schott (1939, p. 413) for the last Wisconsin age, but examination of core P-137 shows no increase in the amount of detrital material more than sufficient to offset a decrease in the abundance of organisms during these ages, and there is no contradiction in these opposing results. This increase in the amount of detrital material was Schott's explanation for the increased rates in the last ice age.

RED CLAY FROM THE PACIFIC OCEAN

The rates of deposition of Red clay (P-259) are given for the upper 160 cm. only. As has been pointed out, the method becomes inaccurate beyond 300,000 years. This core was secured by a different technique (Emery and Dietz, 1939; 1941), and hence the application of the distortion data, obtained with the authors' coring apparatus, is doubtful. There are indications that the rate of deposition has increased in recent times. Certainly the rate of sedimentation of Red clay, as exemplified by this core, is lower than that of the other types of ocean sediments which have been studied, and in this respect the results agree with previous estimates. The mean rate for the past 300,000 years is 0.5 (0.6) cm. per 1000 years, which may be compared with Twenhofel's (1929) estimate of 0.35 cm. of inorganic material deposited on the sea bottom, remote from coasts, in 1000 years. Schott's (1935) estimate for Red clay is <0.86 cm. per 1000 years, with a maximum observed rate of 1.33 cm. and a minimum of <0.5 cm. The rate of deposition of this Red clay, as shown in Figure 4, is 0.25 cm. per 1000 years, 300,000 years ago, or 0.35 cm. if the distortion data be applied. The remaining 86 cm. of core below this point must represent at least 86 cm. of sediment (246 less 160 cm.) and possibly 175 cm., if the distortion data can be applied. Assuming the rates in each case to be those just quoted for the 300,000 year horizon, one should add 350,000 or 500,000 years for the lower portion of the core, to which the present method of dating is inapplicable.

Emery and Dietz (1941) are of the opinion that the ratio of core length to penetration is constant at 0.73 in this Red-clay core. The rates of sedimentation will then be given by the solid line in Figure 4 (P-259) multiplied by the factor 1/0.73. By comparison with the rates of deposition calculated with the distortion data (dashed curve, Fig. 4) these rates are higher near the top of the core but become nearly the same beyond 100,000 years. The mean rate for the past 300,000 years is 0 7 cm. per 1000 years, and the time required to deposit the lower portion of the core remains the same as if no distortion took place—namely, 350,000 years—giving a total time of 650,000 years for the sedimentation of the whole core.

COMPACTION

Apart from the distortion introduced by sampling, there may be compaction of the ocean sediments as they lie in the bottom. The apparently high rate of deposition today may be due in part to compaction, as the sediments become buried. The answer to this problem is given by expressing the results in Figure 4 in terms of milligrams of water-free matter deposited per square centimeter per year. The conversion of units requires a knowledge of the weight of dried material in a unit volume of the original sediment. The measurements of this quantity have been described elsewhere (Piggot and Urry, 1942; Urry and Piggot, 1942).

TABLE 2—*Rates of deposition of ocean sediments in milligrams of water-free matter per square centimeter per year*

Time interval in the past (years)	Zone[a]	Mean rate of deposition in milligrams per square centimeter per year A[b]	B[b]
		P-124 (3) N. Atlantic	
0–1,000	WS	39	41
1,000–1,500	WS	38	43
1,500–2,000	WS	32	36
2,000–4,000	WS	22	27
4,000–10,000	WS	15	25
10,000–12,000	WS	11	32
		P-130 (9) N. Atlantic	
0–1,000	WS	21	21
1,000–3,000	WS	13	15
3,000–4,000	WS	8	10
4,000–12,000	WS	6	7
12,000–20,000	W-G	7	11
20,000–24,000	G	7	19
		P-126 (5) N. Atlantic	
0–8,000	WS	—	—
8,000–12,000	WS	0.9	1.0
12,000–15,000	W	2.2	2.3
15,000–24,000	G	4.3	4.7
24,000–28,000	HW	7.9[c]	9.1[c]
28,000–42,000	HW	2.2	2.8
42,000–51,000	G-W-G	4.2	5.9
51,000–62,000	WS-W-G	3.6	5.8
62,000–73,000	W	3.4	8.4
		P-137 Globigerina ooze	
0–2,500		5.0	5.1
2,500–4,500		3.0	3.2
4,500–8,500		1.5	1.5
8,500–200,000		0.7	0.9
200,000–300,000		0.4	1.2
		P-259 Red Clay	
0–20,000		0.7	0.7
20,000–60,000		0.7	0.7
60,000–160,000		0.4	0.45
160,000–300,000		0.2	0.25

---

[a] This column gives the zones illustrated in Figures 2 and 4 as follows: W, warm-water deposits; S, volcanic shards present; HW, the warm-water? zone of intermediate character; G, glacial marine deposits.

[b] Column A gives the rate of deposition with the assumption that the core length represents the sediment sampled. Column B gives the rate of deposition with the assumption that the relations between core length and the depth of sediment sampled are those found for the varved clays of Connecticut. The true rate of deposition is probably closer to the values given in column B.

[c] These rates do not include the number of centimeters and hence the mass of material in the anomalous zones. (*See* Figure 2.)

DEPOSITION IN TERMS OF MASS

Table 2 summarizes the rates of deposition expressed in milligrams per square centimeter per year. As in Figure 4, two sets of results must be given for the limiting cases of distortion. The relative fluctuations of the rates of deposition are qualitatively similar to those presented in Figure 4, and, in particular, the high rate of deposition at present, as measured in centimeters per 1000 years, is due to an increase in the mass of material precipitated in recent times.

SUMMARY

The ocean floor is covered by deposits of unknown thickness which, for the most part, are stratigraphically undisturbed. Representative samples of several meters of this sediment can now be obtained from below as much as 6000 meters of water. Such core samples provide abundant lithological, chemical, and biological evidence for all the major events occurring within the time that they represent, which in some cases is nearly all of the Pleistocene. Longer cores may eventually penetrate the Pliocene. If these events can be dated, a considerable advance is possible in the chronology of the Pleistocene epoch, for the events occurring not only in the oceans, but also on the continents.

Radium measurements throughout the length of the cores were undertaken in the Geophysical Laboratory, primarily to solve the problem of the high radium concentrations in the surface of some ocean sediments (Piggot, 1933). The solution of this problem provided the basis for a method of measuring time in ocean-core samples. If there are not certain ratios between the numbers of the various types of radioactive atoms incorporated in the ocean sediments, the concentrations of these radioelements will change, with time, in such a manner as to establish these ratios. By following the change of concentration of an appropriate element, this property of radioactivity provides a method of measuring intervals of time within the past 300,000 years. It does not involve the accumulation of either of the stable products of radioactive disintegration—*i.e.*, helium or lead—but requires only a knowledge of the relative concentrations of a radioelement at different depths in the sediment.

The radioelements are introduced into the sediments in amounts differing from the "equilibrium" ratio mentioned above. This accounts for the high radium content of certain deep ocean deposits, compared to similar concentrations in continental formations (Piggot and Urry, 1941). The dates in a core once having been established, a calculation of the rate of deposition should be simple, but, while there is little doubt that the core represents everything that it penetrates, there is no doubt that the length of a particular portion of core is not always the same as the sedi-

ment represented by that sample, particularly near the bottom of the core. This introduces uncertainty into the estimates of rates of deposition derived from previous studies of cores and impairs, to some extent, the present study even though measurements were made to determine the distortion with the same core sampler used to obtain four of the five cores discussed here.

Cores of the sediments of the North Atlantic were studied in great detail by members of the U. S. Geological Survey and their collaborators. Dates are given here for the geological events, in the North Atlantic, discovered in their studies.

Time intervals have been calculated for one core of Globigerina ooze, underlying the Caribbean Sea, which may be applied to Cushman's detailed survey of the Foraminifera in this core.

The radium in a core of Red clay from the Pacific Ocean was determined with the object of throwing some light on the high concentration of radium in the surface of Red clays in general and in order to study the rate of deposition of Red clay.

The rates of deposition in the past, as well as at present, are given for all the core samples studied. The rates given for the upper portions of the cores are probably nearly correct, but one should be cautious in drawing conclusions from changes in the rate in the lower part of a core. However, there is little doubt that deposition is as rapid today as at any time during the period represented by the cores studied here, if not more rapid.

NOTES ON THE CORES OF OCEAN-BOTTOM SEDIMENTS

P–124 Collected on board the cable ship LORD KELVIN May 1936 Station No. 3. Lat. 46° 03′ N, Long. 43° 23′ W. Depth of water 4710 meters Length of core 285 cm.

P–126 Collected on board the cable ship LORD KELVIN May 1936. Station No. 5. Lat. 48° 38′ N, Long 36° 01′ W Depth of water 4760 meters Length of core 282 cm.

P–130 Collected on board the cable ship LORD KELVIN May 1936. Station No. 9. Lat. 49° 40′ N., Long. 28° 29′ W. Depth of water 3750 meters. Length of core 276 cm.

P–137 Collected on board the research ketch ATLANTIS February 1937. Lat 19° 14′ N., Long. 80° 20′ W. Depth of water 4893 meters Length of core 190 cm

P–259 Collected on board the motor vessel E. W. SCRIPPS by F. P. Shepard and R. Revelle with their core sampler. Core No. FPS 186, August 1938. Lat. 30° 41′ N., Long. 121° 46′ W. Depth of water 4080 meters. Length of core 246 cm.

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GLACIATION OF SHICKSHOCK MOUNTAINS, GASPE PENINSULA

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ABSTRACT

The Shickshock Mountains constitute one of the highland areas that have been considered by Coleman on geologic evidence, and by Fernald on botanical and geologic evidence, to have been nunataks during the maximum of Wisconsin glaciation.

Recent field studies have been made on the two highest parts of the Shickshocks—the broad plateaulike masses known as Mount Albert (summit altitude 3775 feet) and Tabletop Mountain (4230 feet). Glacial erratics possibly derived from pre-Cambrian rocks from north of the St. Lawrence were found as high as 3760 feet on Mount Albert. On Tabletop Mountain, striated surfaces were found as high as 3500 feet, and glacial erratics (of local origin) as high as 3700 to 3800 feet Detailed search failed to produce any direct evidence of glaciation through the highest 400-500 feet Throughout this distance evidences of any possible glaciated surfaces are obscured by mantles of locally derived felsenmeer that presumably originated in postglacial time In addition, the composition of the bedrock is so heterogeneous that several types of erratic stones might be present without having been recognized These circumstances do not demonstrate glaciation of the highest parts of Tabletop Mountain; yet there is no geologic evidence inconsistent with the possibility that the entire Shickshock highland was overtopped by Wisconsin ice. No geologic evidence was found suggesting that any portion of the area existed as a nunatak during the Wisconsin maximum.

During the waning stages of glaciation, and possibly earlier, Tabletop was the site of radial outflow from a local ice cap.

INTRODUCTION

The glacial history of the Shickshock Mountains of the Gaspe Peninsula has been variously interpreted, the chief point at issue being whether or not the highest summits existed as nunataks during the maximum of Wisconsin glaciation. Coleman (1922, p. 14, 37) on the basis of geologic evidence and Fernald (1925, p. 295-303) on the basis of both botanical and geologic evidence have favored the nunatak hypothesis. Mailhiot (1919, p. 146) and Alcock (1935, p. 125-126), on the other hand, expressed the opinion that some of the highest summits have been glaciated. As a result of field investigations in August 1941, the present authors are convinced that the existence of nunataks has not been proved. However, they could find no proof that the highest summits were glaciated. It is nevertheless demonstrated that at some time ice reached to within at least 500 feet of the highest summit (Mt. Jacques Cartier) and was at least 1200 feet higher than believed by Coleman. It is thought that this highest ice was of Wisconsin age; it may have been part of the great Labradorian ice sheet or it may have been part of a local ice cap centered over the Tabletop plateau on which Mt. Jacques Cartier is situated.

This study and others in similar highland areas such as the Torngat Mountains in Northern Labrador (Odell, 1938, p. 204-207), Mt. Katahdin in Maine (Tarr, 1900; R. P. Goldthwait, unpublished communication), Mt. Washington (R. P. Goldthwait, 1940, p. 17-19), the Long Range of Newfoundland (MacClintock and Twenhofel, 1940), and the Adirondack Mountains (J. P. Marble, R. Balk, unpublished communications) show that there is no apparent geologic evidence that highland areas in eastern North America existed as nunataks during the maximum of Wisconsin glaciation.

ACKNOWLEDGMENTS

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HISTORICAL SUMMARY

Apparently the first geologist to mention the glacial features of the summit plateaus of the Shickshocks was Mailhiot (1919, p. 146) who held that the summit of Mt. Albert had been glaciated. Coleman (1922, p. 12-14, 31-39), after visiting the summits in 1918 and 1919, concluded that the Tabletop and Mt. Albert plateaus had not been crossed by glacier ice and that Tabletop had not even been covered by a local ice cap. However, he recognized that Mt. Albert might have been a center of radial glaciation. Also, he showed that the Labrador ice sheet crossed Gaspe in the region of the low pass (altitude 750 feet) at the head of the Matapedia River, west of the Shickshocks.

In 1921-1924 Alcock (1926, p. 23-25) studied the geology of the Mt. Albert area and also visited Tabletop and inferred, as Coleman had done, that the summits had not been glaciated. However, after several years' further field study of Gaspe, he (Alcock, 1935, p. 125-126) changed this opinion and concluded that the summit areas probably had been covered by the Labradorian ice sheet.

In 1932 Jones (1933, p. 28) investigated the geology of Tabletop and found no evidence of glaciation at altitudes higher than the heads of the cirques recognized by Coleman.

A botanical party led by M. L. Fernald visited Tabletop and Mt. Albert in 1904. Much later, Fernald (1925, p. 295-303) correlated the distribution of plants in this region with the distribution of glaciated and nonglaciated areas as inferred by Coleman (1922) and later by Alcock (1926).

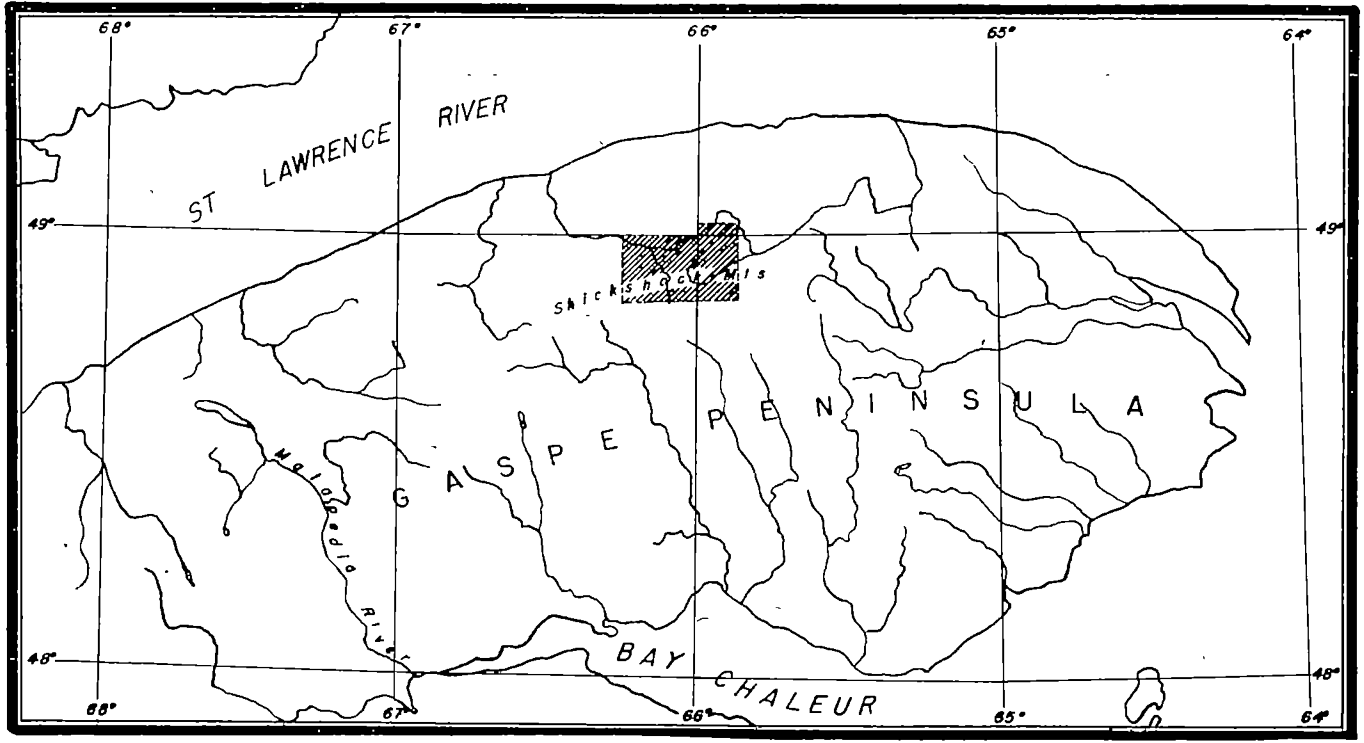

FIGURE 1.—*Index map of Gaspe Peninsula*

Area of Figure 2 is shown by shading.

SHICKSHOCK MOUNTAINS

GENERAL DESCRIPTION[1]

The Shickshock Mountains, lying north of the central part of the Gaspe Peninsula, constitute its highest part. The Shickshock axis is essentially a continuation, via the Notre Dame Mountains of Southeastern Quebec, of the Green Mountains of Vermont. Trending east-northeast, the Shickshocks constitute a belt having a maximum width of 15 miles and lying no more than 25 miles south of the St. Lawrence They consist of deformed Paleozoic sedimentary and volcanic rocks with some large intrusive masses, the whole belt being slightly convex toward the north.

The areas underlain by resistant rocks—the volcanics and intrusives—have broad summits with nearly flat to very irregular surfaces. Alcock (1935, p. 109) relates these areas to an erosion surface that has been warped up and isolated through reduction of the surrounding weak-rock areas during a later cycle of erosion.

The highest parts of the Shickshock Mountains, lying close to the intersection of the 49th Parallel and the 66th Meridian, consist of Tabletop Mountain, whose highest summit reaches 4230 feet, and Mt. Albert, with an extreme altitude of 3775 feet (Figs. 1, 2). Both masses are extensive and are developed on resistant intrusive rocks.

The Shickshock region is densely wooded, chiefly with spruce, fir, and birch, but above altitudes of 2800 to 3300 feet trees (mostly stunted spruce) occur only in patches separated by extensive areas of mosses, grasses, and bare rock.

MOUNT ALBERT

LITHOLOGY

The higher portions of Mount Albert (Fig. 2) consist of serpentinized peridotite weathered buff, encircled by dark amphibolite. (*Cf.* Coleman, 1922, p. 31; Alcock, 1926, p. 34-38.) The serpentinized peridotite characterizes the south peak and all of the plateau except the narrow ridge on the north side, which is composed of the amphibolite.

TOPOGRAPHY

The most striking topographic feature of Mount Albert is the plateau-like character of the higher portions (Pl. 1, fig. 1). The name "tabletop" is much more descriptive of Mount Albert than it is of Tabletop Mountain. The Mount Albert plateau has roughly the shape of a thick-set, irregular letter C with the concavity opening eastward. From north to

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[1] An excellent discussion of the physical features of the Gaspe Peninsula and their origin is given by Alcock (1935, p 109-119).

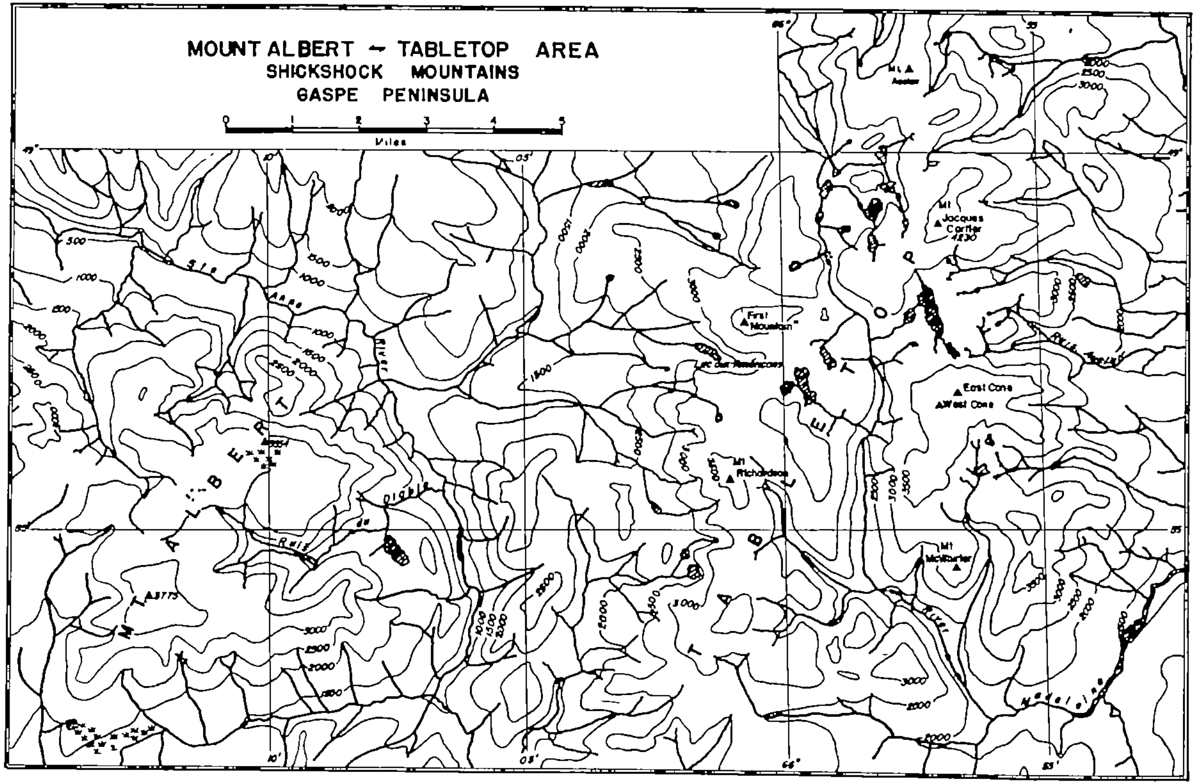

FIGURE 2.—*Map of Mount Albert-Tabletop area*

Compiled from maps of Geological Survey of Canada and Quebec Bureau of Mines (*See* Alcock, 1926, and Jones, 1933.) Because of differences between the maps the present compilation includes adjustments along the 66th Meridian.

south the plateau is about 3 miles long. In an east-west direction the two arms of the C are about 2 and 3½ miles long, the southern arm being the longer. The plateau is bounded on all sides by rather abrupt slopes, some of which descend steeply to the base of the mountain. Cirquelike valley heads indent the plateau in several places, a notable one being on the east side at the head of Rivière du Diable. This valley head is chiefly responsible for the C-like shape of the plateau.

The highest point of Mount Albert is the south peak, with an altitude of 3775 feet. The north peak, forming the highest point on the amphibolite ridge about 3 miles away, has an altitude of 3554 feet and is the more abrupt. From the base of the north peak and from the crest of the south peak, the surface slopes gently and evenly down to about 3300 feet at the lowest and narrowest portion of the plateau, approximately midway between the two peaks. The maximum differential relief on the plateau is thus about 500 feet. A number of small lakes constitute a striking feature of the low, narrow portion. Alpine types of vegetation dominate some areas of the plateau. Other areas are characterized by bare-rock knobs or by felsenmeer—continuous sheets of fragments of frost-disrupted local rocks in process of slow transport down the slopes. Stone-polygons are well developed in places.

EVIDENCE OF GLACIATION

Erratic stones.—Erratics are far from common on the Mount Albert plateau. Possible explanations of their scarcity here and on the other highland areas are discussed in another section of this report. Mailhiot (1919, p. 146) was the first to record the presence of erratics. He says: "On the summit plateau are found a few erratic blocks of granite and hornblende schist which certainly represent remnants of ancient glacial moraines." According to Coleman (1922, p. 31), "The mountain is really a steeply walled tableland covered with weathered masses of serpentine except for a ridge of garnetiferous hornblende schist which rises a little higher than the rest and has rolled fragments down on the serpentine." Coleman's explanation of the occurrence of amphibolite on serpentinized peridotite may be correct for the immediate vicinity of the ridge, but this explanation does not seem possible for an amphibolite erratic which one of the present authors found within 100 feet of the summit of the south peak and thus about 100 feet higher than the summit of the amphibolite ridge 3 miles away.

Mailhiot (1919, p. 146) mentions granites on the plateau, *i.e.*, at altitudes of more than 3300 feet. Coleman (1922, p. 32) mentions granite boulders up to about 1500 feet only. According to an unpublished communication from F. J. Alcock, a gold-quartz boulder of probable Labradorian origin was recently found at an altitude of about 1550 feet, about 6 miles south of the summit of Mount Albert.[2] Alcock (1926, p. 24) found one 10 feet in diameter at an altitude of 2500 feet and states that two others occur at a similar altitude. Alcock also records a conglomerate boulder about 20 feet long at an altitude of about 3200 feet.

The present investigation disclosed erratics at still higher altitudes. Quartzites were found on the plateau at altitudes of over 3300 feet; a granite-gneiss cobble was collected within 50 feet of the summit of the south peak, and a granite pebble within 15 feet of the same summit. Stones of serpentinized peridotite were observed within 15 feet of the summit of the north peak. According to Alcock (unpublished communication) the highest erratic and several others found at lower altitudes (biotitic quartzite or paragneiss at about 3450 feet; garnetiferous sericite schist at about 3100 feet) are probably of pre-Cambrian origin from across the St. Lawrence and hence may have been transported by Labradorian ice. However, as pointed out by Alcock and emphasized by Jones (unpublished communication), the erratics are not sufficiently distinctive to prove such an origin.

That these erratics could not have been deposited by streams on the preglacial erosion surface represented by the plateaus of Mount Albert and Tabletop and other upland areas (Alcock, 1935, p. 109) is demonstrated by the following evidence: (1) The granitic rocks observed on the south peak are firm and but little weathered and could hardly have been subjected to weathering throughout Pleistocene time; in fact their degree of weathering is consistent with a Wisconsin age; (2) the quartzites and conglomerate rocks must have been derived from lower places; (3) several of the erratics are too large to have been transported by streams on an erosion surface of peneplanelike character; (4) the serpentinized peridotite boulders on the north peak can hardly be regarded as derived from the higher serpentinized peridotite area by mass-wasting or fluvial transport. The present topography would not permit this, and, if topographic reversal has occurred in the vicinity of the north peak since their deposition, the serpentinized peridotite erratics would have had to survive while a considerable prism of bedrock of the same lithology was completely removed. Thus, with the possible exception of the amphibolite stones lying on serpentinized peridotite at the base of the north peak, all the erratics must have been glacially transported. It therefore follows that ice covered Mount Albert up to within at least 15 feet of the top of the highest peak.

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[2] The boulder, 3½ feet in diameter, was found by Mr. Beidelman, of the Federal Zinc and Lead Company, about 100 feet above Brandy Brook on its northwest side, near the northwest corner of mining claim 1563 (Cf Alcock, 1928, Map 2129, 210A)

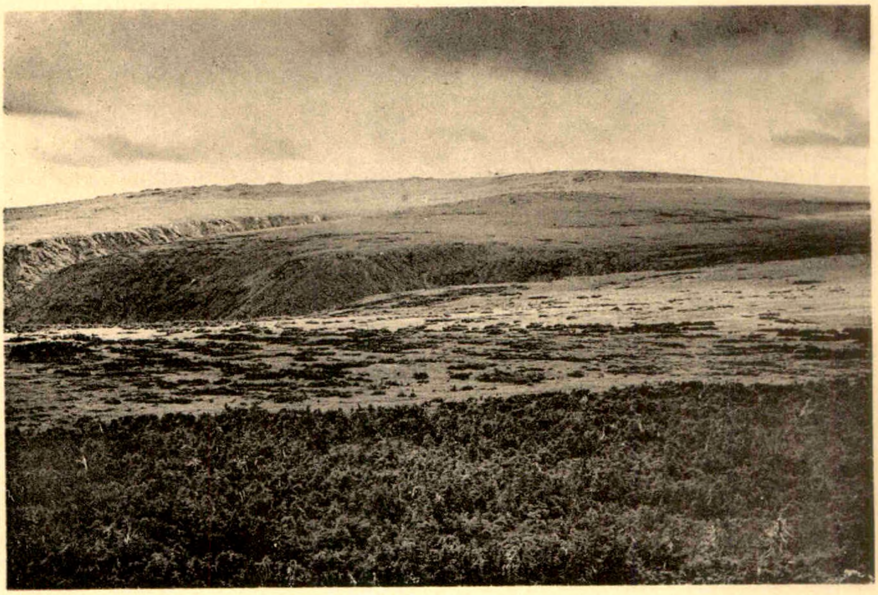

FIGURE 1. MOUNT ALBERT PLATEAU
View from north peak toward south peak.

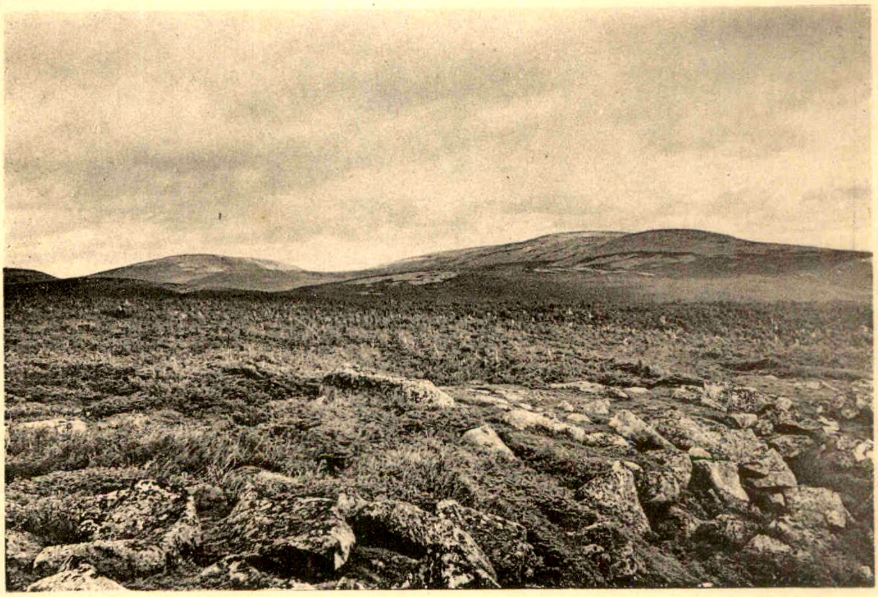

FIGURE 2. PLATEAULIKE SURFACE OF TABLETOP
View from near "First Mountain" northeast toward Mount Jacques Cartier.

SUMMITS OF MOUNT ALBERT AND TABLETOP

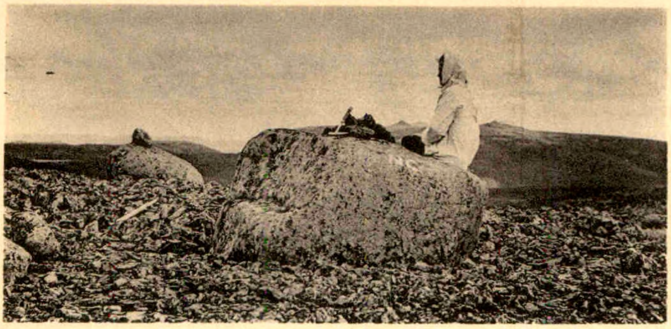

FIGURE 1.
LARGE
GRANITE
ERRATIC
On summit
of "First
Mountain".

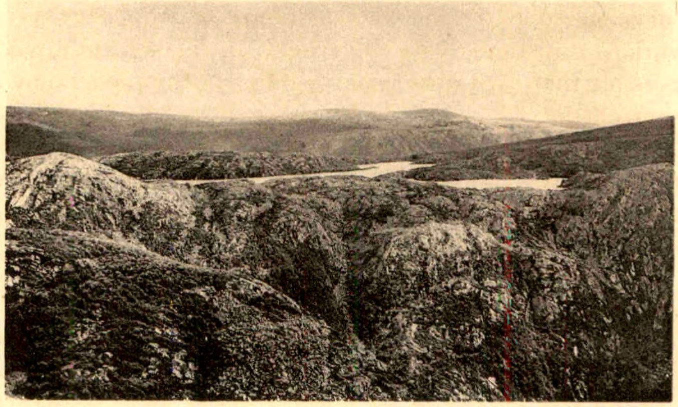

FIGURE 2.
ROCK-BASIN
LAKES
Steep slope
in foreground
is head of
Lac aux
Américains
cirque.

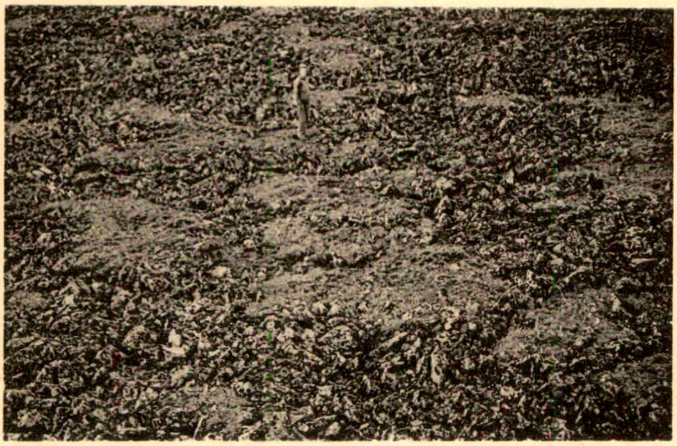

FIGURE 3.
FELSENMEER
WITH STONE-
POLYGONS
Summit of
Mount
Jacques
Cartier.

VIEWS ON TABLETOP

Glaciated surfaces.—A number of irregular rock-basin lakes, several yards to several tens of yards in diameter, occur in the low, narrow portion of the plateau at about 3300 feet. Being in serpentinized peridotite, the basins cannot be of solutional origin, and the writers know of no other process except glaciation which would form basins of this character. Moreover, the fact that even very small rock-basin lakes are not filled with sediment argues for recency of glaciation.

No other definite products of glaciation were noted on the Mount Albert plateau. In several places there is a suggestion of stoss-and-lee topography which could be interpreted to indicate ice movement from the northwest across the plateau (*cf.* Mailhiot, 1919, p. 146); a tendency for the lakes to be elongate in a northwest-southeast direction could be taken as support for such an interpretation. However, as jointing in various directions is well developed in the serpentinized peridotite, it is possible that such directional features in the topography are structurally controlled.

The fact that no striae or other glacial markings were observed by Coleman, Alcock, and the present writers is not to be wondered at, for the surfaces of the bare-rock areas are so much weathered that they could not be expected to retain such features.

CHARACTER OF MOUNT ALBERT GLACIATION

Coleman (1922, p. 31) grants that ". . . ice may have moved outwards from it [Mount Albert]." According to Alcock (1926, p. 24),

> "Local glaciers occupied the flanks of Mount Albert during the Pleistocene. On the west and south sides of the mountain are a number of amphitheatres which are undoubtedly cirques formed by small glaciers. A rather striking amphitheatre lies about a mile west of the north geodetic station. It is well defined, with steep walls 500 feet high and a flat bottom in which lies a small lake Other steep-walled cirques to the south have a depth of 1,000 feet."

Later Alcock (1935, p. 124-126), on the basis of indirect evidence, suggested that, in addition to being the site of local glaciation, the high Shickshocks, including Mount Albert, were crossed by Labradorian ice, thus abandoning his earlier (1926) view that this had not occurred.

Observations of the present authors prove that the summit plateau was also glaciated. The fact that rocks foreign to the plateau occur among the erratics shows that the ice which covered the plateau originated elsewhere. However, the lithology of the erratics does not prove whence the ice came. As previously pointed out some of these erratics may have originated in localities north of the St. Lawrence River, in which case they were brought to the Mount Albert region by Labradorian ice. However, even if such Labradorian origin were demonstrable, there would still be no proof that Labradorian ice ever covered the summit of Mount

Albert. This is because there is a possibility that Labradorian ice may have deposited erratics at a low altitude in valleys between Mount Albert and Tabletop and that subsequent redistribution by radial outflow from an ice cap on the Tabletop plateau brought them to the summit area of Mount Albert.

Thus it appears that Mount Albert has been the site of two kinds of glaciation: (1) General glaciation that covered Mount Albert. This was either continental (Labradorian), or local (originating on Tabletop), or both; (2) strictly local glaciation; certainly cirque glaciation, perhaps preceded by a small Mount Albert ice cap.

TABLETOP MOUNTAIN

LITHOLOGY

Tabletop Mountain (Pl. 1) consists of a granitic intrusive about 10 miles long by 5 miles wide and of irregular form. The rock varieties seen by the writers consist of granite, syenite, granodiorite, diorite, granite porphyry, felsite, and pegmatite. These have been described by Jones (1933, p. 24-25) who believes that they are differentiates of a single magma. Surrounding the intrusive is a rather narrow zone of somewhat altered rocks, consisting chiefly of slates altered to hornfels and calcareous sediments altered to lime-silicate rocks.

TOPOGRAPHY

Tabletop Mountain is a kind of dissected plateau, although the surface is not what is ordinarily thought of as plateaulike, inasmuch as the maximum relief of this small area is nearly 1000 feet, exclusive of the capacious valley of the Madeleine River that is incised into the southern half of its central part.

The outer flanks of the plateau are very steep slopes, in places 2000 feet high. These are indented by deep valleys with cirquelike heads, the altitudes of the floors averaging a little more than 2500 feet. The plateau has a high irregular rim held up by resistant rocks: chiefly metamorphics on the west side and chiefly granites and related rocks on the east side. The main area inclosed within this rim is several hundred feet lower than the rim itself and is occupied by a number of lakes.

The domelike peaks that together constitute the rim are nearly bare of vegetation and are extensively covered with felsenmeer. In places this superficial debris is organized into stone-polygons and stone-stripes. The highest of these peaks is Mt. Jacques Cartier with an altitude of 4230 feet (Pl. 1, fig. 2); others are Mt. Auclair, Mt. Richardson, Mt. McWhirter, and the broad ridge that bears East Cone and West Cone.

EVIDENCE OF GLACIATION

Glacial striae.—Although Coleman (1922, p. 14) found no striae at altitudes greater than 2500 feet, the writers found striae at two localities at much higher altitudes. One locality (Fig. 1) is at an altitude of about 2900 feet on the trail ascending the west flank of Tabletop Mountain from Lac aux Américains. The striae occur on a smooth boss of lime-silicate rock that has a surface sloping in a direction S. 12° W. at about 50° from the horizontal. On this surface the striae trend S. 83° W. and slope down toward the southwest at 18° from the horizontal. One hundred feet higher on this same trail is another boss of similar rock with striae trending S. 82° E.

In the writers' opinion these two occurrences of striae are of glacial origin and therefore demonstrate glaciation at this place. Unfortunately the direction of ice flow cannot be surely inferred from them so that no discrimination could be made between the possibility of ice encroaching upon the Shickshock highland from the region north of the St. Lawrence and the possibility of ice flowing radially outward from the Shickshock highland itself.

The other locality at which striae occur is on the north flank of the peak called by Coleman "First Mountain," nearly a mile northeast of the first locality and at an altitude of about 3500 feet. The rock here is hornfelslike. One boss has a face sloping north at about 45° from the horizontal; on this face are striae believed to be glacial, trending obliquely to it, but no features were observed that would indicate the direction of ice flow.

All that can surely be inferred from the observed striae is that glaciation on the west side of Tabletop Mountain up to an altitude of at least 3500 feet is demonstrated. Although a search was made for other striae, none was found. This is explained at least in part by the fact that the low central area of the plateau is covered with vegetation and the barren eastern part consists largely of granite whose surfaces, roughened by weathering at these altitudes, are not capable of preserving striae. In any case striae on the east side of Tabletop would not discriminate between Labradorian ice and local ice, for either would have flowed in the same direction, namely, southeast.

Erratic stones.—On the west rim of Tabletop, stones and boulders of Tabletop granite rest on the flanks and summit of "First Mountain," a peak consisting entirely of metamorphic rocks. These stones mentioned by both Coleman (1922, p. 37) and Alcock (1926, p. 24) occur by the hundreds, and one of them, an angular block (Pl. 2, fig. 1) 10 by 7 by 3½ feet, lies on the very summit, at an altitude of between 3700 and 3800 feet. The highest of these stones have been transported horizontally not less than 1000 feet toward the west and in the process have been lifted through a minimum of 100 feet. Coleman attempted to explain such occurrences as the result of topographic reversal. He believed that the immediately adjacent areas of granite on the east, which now lie about 100 feet lower than the summit of "First Mountain," formerly stood so high that great blocks of granite were shed downslope upon the lower metamorphic area that now constitutes "First Mountain." According to this hypothesis, the present erratics are mere residual weathering-cores of these granite blocks. To the writers this hypothesis is incredible because it demands that a higher area of granite should have been eroded to its present altitude, lower than metamorphic rocks of "First Mountain," during a time when isolated erratic blocks of granite have withstood long continued weathering.

Nor are these erratic granites the products of stream transport during the development of the preglacial erosion surface inferred by Alcock, because (1) many of them are too large to have been transported by this means, and (2) granitic rock can hardly be expected to have survived in firm fresh condition throughout Pleistocene time.

Hence the opinion is reached that these granitic boulders are glacial erratics and that as such they demonstrate glaciation up to at least 3700 feet (more than 200 feet higher than the highest striae seen). Further, they demonstrate movement westward (outward) radially away from Tabletop Mountain.

On other parts of Tabletop there are also many boulders and small stones of granitic and certain metamorphic rocks. Most of these are residual. On the other hand some of them may have been transported by glacier ice to their present positions, but they are so like the assemblage of local rocks that constitute the felsenmeer in which they occur that derivation as erratics cannot be proved.

The scarcity of stones of proved glacial origin on Tabletop is characteristic of the whole Gaspe Peninsula, as has been recognized by all geologists who have worked in Gaspe. This is partly because the drift, however scanty, includes a very large proportion of local rock types and partly because the dense cover of timber makes observation difficult, and stones of any kind are not easily seen except along valley floors where they have been reworked by streams. Other factors involved are discussed in a later part of this paper.

Broad glaciated surfaces.—Tabletop is marked in places both by rock basins and by broad smoothed undulating bedrock surfaces truncating joint structures. Neither of these features occurs on the highest summits, where only felsenmeer is in evidence, but both occur at various places elsewhere on Tabletop. Examples of broad glaciated surfaces are found east of the head of the Lac aux Américains cirque, at 3500 feet, and west of the head of the cirquelike valley of Ruisseau Auclair, at about 3300 feet. The lakes in the former locality have rock basins (Pl. 2, fig. 2), as have those one mile to the north (Coleman, 1922, p. 36). Some of the lakes 1 to 2 miles northwest of Mt. Jacques Cartier are rimmed by vegetation and may not lie in individual rock basins.

These surfaces exhibit no glacial polish in detail, for they have been roughened by weathering, yet the writers believe they are good evidence of glaciation.

Close relation of mantle to underlying rock.—In summit areas where the relation of the frost-weathered mantle (felsenmeer) to the underlying rocks can be seen, the lithologic character of the mantle changes rather abruptly at the contact between any two kinds of bedrock. If the summits had been undergoing weathering and mass-wasting since preglacial time the resulting downslope movements of the mantle should have distributed debris far downslope across geologic contacts. As this does not appear to have occurred, it may be reasoned that these processes have been in operation only a short time, prior to which any previous accumulation had been removed. This relationship constitutes strong indirect evidence that the preglacial mantle was removed by glaciation.

Summary.—The direct evidence of glaciation on Tabletop consists of broad glaciated rock surfaces, including rock basins, as high as 3500 feet, glacial striae as high as 3500 feet, and erratic stones at more than 3700 feet. Immediate indirect evidence consists of a close relation of coarse, frost-weathered mantle to the immediately underlying bedrock. Only three summits reach or exceed 4000 feet (Mt. Jacques Cartier and the unnamed summits immediately north and southwest of it, respectively). From these no direct evidence of glaciation has been obtained.[8]

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[8] In discussing Mt. Washington, New Hampshire, and Mt. Katahdin, Maine, both of which were submerged beneath the last ice sheet, Antevs (1932, p. 7, 8) stresses the absence of a recognizable upper limit of glaciation as evidence that these peaks were glaciated. The glacial features, distinct on lower slopes, become less and less distinct with increasing altitude. However, this fact, although suggestive, is hardly evidence of glaciation of the summits. In the mountainous highland of northeastern Washington, where, according to observations made by the senior author, all gradations occur from summits once wholly covered by the last glacier ice to extensive nunataks, the upper limit of former glaciation is hardly anywhere distinct but has to be inferred from the altitudes of the highest foreign stones on the mountainsides. The authors see no reason why an ice sheet in a mountainous region should necessarily leave a well-defined record of the highest position of its surface.

CASE FOR NONGLACIATION

Coleman's (1922) conclusions on the glaciation of the high Shickshocks were as follows:

> "There is no evidence that glaciers worked on the highest parts of the Shickshock mountains, for their surfaces consist of loose blocks of the underlying rock, seldom showing any displacement; and no striated surfaces or boulder clay have been found above 2 500 feet" (p 14)
>
> "It [Mount Albert] resembles some tablelands in the Torngat mountains of Labrador and seems never to have been crossed by ice, though ice may have moved outwards from it." (p. 31)
>
> "The block-covered summits of the hills and mountains of Tabletop and the entire absence of boulder clay, moraines, or smoothed and striated surfaces argue strongly against an outward movement of ice from the centre; and these features, along with the sharp-edged, northwestward-facing cliffs at the edge of the table, provide conclusive evidence against the supposition that the Labrador ice-sheet ever crossed the Shickshocks." (p 37.)

The writers do not agree with these statements, in part as to fact and in part as to interpretation. With regard to Mount Albert, the presence on its plateau surface of stones derived from lower altitudes proves that it was glaciated, though the possibility is not excluded that the glacier was a local ice cap centered over Tabletop. In the writers' opinion Tabletop itself bears direct evidence of glaciation up to an altitude of at least 3700 feet. Of the several points on Tabletop that stand higher than this, only two were examined, owing to unfavorable weather. Mt. Jacques Cartier, the highest point of all, was examined in great detail. Its broad domelike surface is covered with local rock fragments arranged in polygons at its summit and in stone-stripes and broad felsenmeer on its flanks, and despite intensive search no stones recognizably foreign to the granitic bedrocks of this dome were found upon it.

Referring back to the statements quoted from Coleman, we find that his opinion that the high Shickshocks were not glaciated was based on (1) lack of glacial features at altitudes of more than 2500 feet, and (2) the presence of felsenmeer composed of local rocks in the summit areas. It has been shown that glacial features do occur on the summit of Mount Albert, and on Tabletop up to an altitude of at least 3700 feet. This leaves in question only the 500-foot zone between the highest glacial features and the highest summit (Mt. Jacques Cartier). This zone is indeed largely covered by felsenmeer composed of local rocks (Pl. 2, fig. 3) but, as pointed out by Alcock (1935, p. 125) and believed by the present writers, this is not evidence against glaciation because similar felsenmeer areas occur in places where glaciation has been proved by direct evidence.

One of these is Mt. Washington (6288 feet), dominating the Presidential Range in New Hampshire. Mt. Washington has been held by C. H. Hitchcock (1904, p. 83), J. W. Goldthwait (1913), Antevs (1932, p.

1-7), and R. P. Goldthwait (1940, p. 17-19) on a basis of rare though unmistakable striae, till, and erratic stones, to have been overtopped by glacier ice within late Pleistocene time, and this despite the fact that extensive felsenmeer, stone-stripes, and stone-polygons have developed on its summit area since the mountain emerged from beneath the ice. Upon leaving the Shickshocks, the writers visited Mt. Washington in order to make direct comparisons. The similarity, both as to glacial features where they occur and as to felsenmeer, is striking. The features on Mt. Washington prove that extensive frost wedging and mass-wasting have occurred there since the last glaciation.

However, Mt. Washington is not the only place where extensive frost wedging has produced abundant felsenmeer within postglacial time. The top of Mt. Katahdin in Maine is extensively covered with felsenmeer, yet Tarr (1900, p. 436) found foreign stones on the highest summit (5267 feet) and thereby proved it had been submerged beneath an ice sheet. R. P. Goldthwait (unpublished communication) cites additional proof of the same character. Tarr recognized that, even were the foreign stones lacking, ruggedness of the summit would not prove nonglaciation. His argument was threefold: (1) The mountain projected high into the glacier, well above the zone of maximum glacial erosion. (2) The mountain was a nunatak for a time during deglaciation; this time must be added to the postglacial time available for frost-splitting action. (3) High altitude and position above timber line provide optimum conditions for intensive postglacial sharpening of the summit by frost.

Again, the high summits of the Long Range of Newfoundland bear abundant evidence of glaciation (MacClintock and Twenhofel, 1940, p. 1732-1739) despite the fact that Coleman (1926) considered the summits to record ancient glaciation or none at all. The Gros Morne (2651 feet), one of the highest summits in Newfoundland, was climbed in 1939 by the senior author of the present paper. Its broad domelike surface is completely mantled with a felsenmeer of sharp, ragged fragments of the local quartzite; yet scattered through the felsenmeer are scores of gneisses, dark igneous rocks, and other foreign stones and boulders, one of them, at the very summit, being 4 feet in diameter. The felsenmeer could have been developed only since glaciation.

The mountains of northern Labrador exhibit similar features. In the Torngat region, where the highest peaks slightly exceed 5000 feet, Bell (1884, p. 14DD) found that "the mountains around Nachvak are steep, rough sided, peaked and serrated, and have no appearance of having been glaciated, excepting close to sea-level." Daly (1902, p. 247-251) observed no drift higher than 2050 feet and suspected that no glaciation had occurred at more than 2100 feet. Coleman (1921, p. 26), on similar evidence, went further and stated that the northeastern peninsula of Labrador had not been covered by an ice sheet. More recently Odell (1938, p. 205) found ice-abraded bedrock high on several mountains in the Torngat Range, the highest (perhaps not latest Wisconsin in date) being at an altitude of 4700 feet. Odell does not state whether these occurrences can be explained by radial flow of former local glacier ice or whether they demand the Labradorian ice sheet itself. In Odell's opinion the presence of frost-sharpened peaks and an extensive felsenmeer is not inconsistent with the concept of former ice-sheet glaciation because such features are present in parts of East Greenland and Spitsbergen that were certainly glaciated in late glacial or even later time. Similar conditions are known in Norway (Werenskiold, 1915, p. 359-360; Fig. 2).

These facts and opinions make it clear that postglacial frost splitting and mass-wasting in favorable localities have led to the development of extensive felsenmeer. The writers know of no reason why, analogously, this process should not have operated on the Shickshock summits during the same bracket of time. So far as the writers can learn, no argument supporting a contrary view has appeared in the literature. Until such argument has been advanced, the presence of felsenmeer on the Shickshock summits cannot be accepted as evidence that the area between 3700 feet and 4230 feet—the highest point—has not been glaciated.

DATE OF GLACIATION

The time elapsed since the glaciation of any district is estimated ordinarily by the amount of chemical decomposition of the drift or of glaciated bedrock, and by the estimated probable rate at which decomposition takes place.

The basis for estimating the date of glaciation of the high Shickshocks is not very satisfactory. It may be stated that:

(1) Striated surfaces on lime-silicate rocks are not notably dimmed by weathering processes.

(2) Mechanical weathering (chiefly frost wedging) is strongly in evidence on the summits.

(3) Recognizable chemical decomposition is confined to faint limonitization of some angular blocks of dark rocks to a maximum depth of about 1/8 inch.

(4) Some boulders of coarse-grained granite have lost their corners and have been reduced to a subrounded shape.

(5) The centers of stone-polygons on the summit of Mt. Jacques Cartier consist of clayey silt and stones to a depth in one place of at

least 5 feet. This material is thought to be a product of decomposition of the underlying granite.

(6) Frost wedging is believed to be rapid at this latitude and altitude. The frequency with which the temperature fluctuates through the freezing point is probably great.

(7) Fog and mist are prevalent on the higher parts of Mount Albert and Tabletop during the summer. (6) and (7) are believed to be an adequate explanation of the observed disintegration and decomposition of the granite.

(8) Many rock-basin lakes are not yet filled with peat or with clastic sediment.

In view of these statements the writers believe that the glaciation of the high Shickshocks could have occurred in Wisconsin time and that no evidence is at hand to support the belief that the glaciation dates from pre-Wisconsin time.

SOURCE OF GLACIER

Coleman, Alcock, Jones, and the present writers agree that local glaciers radiated from Tabletop and Mount Albert in late glacial time. This is shown by low-level cirques in the outer slopes of these two plateaulike masses and by evidence of the radial transport of drift from them. The altitude of the upper surface of this radially flowing ice cannot be inferred from the evidence collected. At one time the ice may have been an extensive cap completely burying the high Shickshocks. At its maximum it must have been at least high enough to overwhelm "First Mountain" on the west side of Tabletop, the broad central area of the plateau thus being covered. Above such an ice cap Mt. Jacques Cartier could have stood, if at all, no more than 400 to 500 feet. At a later time it may have thinned and become segmented into several independent glaciers. That this radial flow postdated any invasion by ice from north of the St. Lawrence is suggested by a morainelike feature at Lac aux Américains, whose form, convex downvalley, points to its construction by a glacier flowing westward from Tabletop.

That the high Shickshocks were overtopped by the Labradorian ice sheet cannot be positively affirmed from direct evidence, which must consist of the presence of stones derived from the region of northern Quebec, resting on the highest parts of the Shickshocks. However, the following lines of indirect evidence have been advanced for believing that the Labradorian ice sheet did actually overwhelm the Shickshocks:

(1) According to Alcock (1935, p. 125),

> "The whole plateau country with elevations up to 1,800 feet shows drift, and the Shickshock region appears to be too small to have supplied the drift that spread

> out so far. The evidence of Labradorian erratics [*cf.* Jones and McGerrigle, 1939, p 31] also appears to demand that the ice sheet crossed the peninsula. It either did this or lapped around the Shickshocks"

(2) Also according to Alcock (1935, p. 125),

> "A fact that also must always be considered in connexion with the subject of glaciation in the Shickshocks is that mountains considerably higher than any in that range and much farther away from the center of ice dispersal show evidence of having been crossed by the continental glacier. Mount Katahdin in Maine, elevation 5,150 feet, and Mount Washington in the Presidential range of the White Mountains of New Hampshire, elevation 6,293 feet, both show glaciated surfaces and numerous foreign erratics consisting of granites, gneisses, etc. With a continental glacier crossing these peaks it would seem that a thickness of ice of some 2,000 or 3,000 feet must at that time have covered the highest Shickshocks."

At first sight this argument seems convincing. However, the Shickshocks do not lie in a line connecting the White Mountains or Mt. Katahdin areas with the supposed centers of Labradorian ice dispersal; rather, the Shickshocks are on a radius removed about 20° from a radius through Mt. Katahdin. Thus thick ice could have overwhelmed Katahdin when relatively thin ice existed along the radius toward the Shickshocks. Such a condition might have been brought about because the St. Lawrence lowland, adjacent to the Shickshocks, would have allowed easier and more rapid discharge of ice toward the eastern margin of the ice sheet than toward the south. The present writers do not argue that this must have been the case. The suggestion is made merely to show that Labradorian glaciation of Mt. Katahdin and the White Mountains does not necessarily prove that the Shickshocks were ever overwhelmed by Labradorian ice.

(3) The fact that erratics of possible Labradorian origin occur on the summit plateau of Mount Albert might be taken as evidence that Labradorian ice had crossed this area. However, as already pointed out this evidence is equivocal.

As indicated, these three lines of indirect evidence do not demonstrate that Labradorian ice crossed the Shickshocks, although they are consistent with the possibility that this occurred.

The chief reason why indirect evidence has to be relied on is the rapid upward diminution of drift in the high Shickshocks. This is not peculiar to the Shickshocks but is characteristic of many glaciated regions of great relief and steep slopes. Explanation probably lies in four factors, two of which are inherent in the original distribution of the drift, whereas the other two are inherent in secondary redistribution:

A. Original deposition.

(1) There is an upward component of basal flow in ice that encounters a steep adverse slope. This retards basal flow and thus tends to cause

deposition near the base of the slope, leaving little drift to be deposited at higher positions.

(2) The greater velocity of ice flow in valleys as compared with adjacent intervalley areas tends to concentrate drift-laden ice in the valleys. As a result, when deglaciation occurs there is greater deposition of drift in the valleys than on the intervalley areas.

B. Secondary redistribution.

(3) It is clear, as pointed out in the present paper, that in late glacial time the high Shickshocks constituted a center of radial glacial outflow. This movement must have carried outward and downward a large part of whatever drift had been left at high levels by any pre-existing larger ice sheet. This factor could operate only in those highland areas that were subject to local nourishment and outflow of glacier ice during or following general regional deglaciation.

(4) Mass-wasting and stream erosion following deglaciation has been a powerful factor in many places, tending to emphasize the originally very unequal vertical distribution of the drift. On many steep slopes mass-wasting processes are believed to be competent to remove all of the drift deposited on them during glaciation.

Radial outflow of ice from Tabletop has already been shown to have occurred in late glacial time. That this radial outflow was not confined to the cirque-headed valleys that cut the outer slopes of the Tabletop plateau but occurred on the plateau summit up to an altitude of more than 3700 feet is shown by the granite erratics derived from the central rock mass of Tabletop but now resting on the metamorphic rocks of the western margin. The extent and thickness of the Labradorian ice sheet when this radial flow occurred is not known, except that it must have been thick enough to allow a local gradient, sufficiently steep to enable ice to flow radially away from Tabletop.

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MISSISSIPPI DELTA—A STUDY IN ISOSTASY

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ABSTRACT

The work of Barton, Russell, and the Geological Survey of Louisiana has shown that the Mississippi delta is over 30,000 feet thick and that it has subsided *pari passu* with the deposit of sediment. They have considered isostasy as an explanation of this subsidence, and Barton has rejected the explanation. Russell and others have not satisfied themselves that the subsidence and the great thickness of the delta are due to the maintenance of an isostatic balance which was disturbed by sedimentary loading. The reason for the failure to establish the operation of the principle of isostasy is due to the fact that the load of water displaced by the delta has been ignored as a primary factor in the problem. No cognizance has been taken of the depth of the sea into which the delta was built. In the present paper it is shown that a delta like that of the Mississippi built out into deep water will subside under sedimentary load and will maintain isostatic balance throughout its growth, up to a limit of thickness which is determined by the initial depth of water. Russell's view that the alternation of degradation and aggradation of the surface of the delta is due to the fall and rise of sea level caused by glaciation and deglaciation in Pleistocene time is accepted as proven.

INTRODUCTION

Geological processes characteristically recur in cycles. Every cycle has a beginning and an end, but the process, of which it is a manifestation, never ends if we consider the earth as a whole. The erosional cycle is typical of geological cycles in general. Continental erosion has been active since the beginning of recorded time and is still going on at a vigorous rate. In any cycle of this prolonged process its early stages pertain to the physiographically youthful aspect of the continental surface, when the mountains are new and the relief is high. Then erosion is at its maximum vigor. When the divides become low and the relief, within the hydrographic basin of a master stream, is reduced to so subdued a surface that the running water ceases to function as an agency of erosion, then the cycle is in its closing stage. It can be renewed only by a rejuvenation of the relief due to a recurrence of deformation of the earth's crust or to a lowering of sea level relatively to the continental surface.

Corresponding to the erosional cycle there is a depositional cycle in which, for every drainage system, the waste products of degradation are carried to the sea and there dumped in the form of a great embankment known as a delta. But the two cycles do not necessarily begin or end at the same time, at least insofar as we recognize in the delta the result and measure of deposition. Long before erosion in any hydrographic basin has completed its task and the surface of the land has been reduced to a peneplain, the delta may have reached the limits of growth. Those limits may be set by the strength of the earth's crust, if the sea in which the delta is initiated be deep enough. A delta is a load imposed locally upon the surface of the earth, which as it becomes *large* is accommodated by depression. The sinking is slow, and the upbuilding of the delta at its surface may, and usually does, keep pace with the lowering of its bottom. Thus the embankment, as it advances seaward, takes on the configuration of a slab, or formation, of sediment, resting on the immediately antecedent sea floor. The growth of the delta, or the increase of its dimensions, is thus downward as well as upward and seaward. The vertical dimension at any stage of its growth is the direct measure of the load imposed by the sedimentary process. From this must be deducted the load of sea water displaced. The top of the delta may be partly a land surface and partly subaqueous, and during growth this top is much more stable as to position than is the bottom. The latter slowly subsides under load stress, while the former, particularly as to its subaerial portion, is held relatively fixed by the sea level.

To accommodate the sinking of the embankment into the earth's crust there must be a flowage of rock in depth, out from beneath the depression, equal in mass, per unit of area, to the load imposed. This flowage is mostly landward, whence the load is derived, and tends to balance the diminishing weight of the crust in that direction. When the depression under the delta becomes broad and the embankment of weak incoherent sediments, to a thickness of several miles, becomes an integral part of the crust, the latter enters upon a critical stage of the whole geologic process. The tangentially compressive stress hitherto dormant in the outer part of the earth becomes active, and the weakened crust, together with the contents of the depression, collapses in the form of a geosyncline. The strata of the great embankment are folded, and a mountain range is projected slowly and fitfully far above sea level. The depositional cycle has closed, and another has been inaugurated by the diversion of the river mouth to some other part of the continental coast.

If, however, the load stress, measured in terms of thickness of the delta, together with the corresponding thinning of the strong crust, be insufficient to induce this horizontal collapse, the embankment may grow seaward indefinitely and so greatly prolong the duration of the depositional cycle. As will appear in the sequel the limiting thickness of the delta is determined by the depth of the sea at the inauguration of delta construction. Thus only deltas built out into relatively deep seas develop the *loci* of orogenic axes.

It is the purpose of this paper to review somewhat more in detail the history of such a depositional cycle, making particular reference to the delta of the Mississippi River.

The delta is considered to have been controlled in its growth and to have had its thickness determined by the operation of the principle of isostasy. The effort of the paper is not so much to give a detailed account of the delta as to use it, notwithstanding some defects of data concerning it, as an illustration of how the isostatic mechanism works. In the discussion it is necessary either to make certain assumptions or to treat the problem algebraically. I have preferred to make the assumptions in order to present a concrete picture of this great geological process, feeling confident that other values, which might reasonably have been assumed, would not have changed the picture essentially, but would have resulted in merely numerical differences. The discussion throughout is deductive, and the fundamental assumption is that of the validity of the doctrine of isostasy.

STRUCTURE

Our knowledge of the structure of the great deltas now functioning in various parts of the world has come from a study of small embankments at the mouths of rivers discharging into lake basins. On the shores of

Lake Superior one may stand at the mouths of small streams and watch the process of delta building. The drainage of the hinterland of the lake laden with sand and gravel issues with great velocity from small gorges. The moment that the stream emerges from the confinement of the gorge, the water spreads, becomes shallow, loses velocity, and drops parts of its load. From this point forward to the lake shore the stream has a much lower gradient, across the surface of the embankment which it has built up. Not only is the gradient and therefore the velocity diminished, but the stream divides into several distributaries, which flow radially from near the mouth of the gorge to the curving front of the delta at the shore. These distributaries drop their load of detritus progressively over the top of the embankment as they flow to the lake. They also bifurcate and otherwise subdivide, and the position of their channels is very inconstant. They thus spread the stream detritus fairly evenly over the top surface of the delta. The only unevenness worthy of note is due to the fact that the maximum dropping of stream load is at the borders of the channels, so that the latter become situated on low ridges. It is this precarious perch of the channels which induces their frequent change of course. At the brink of the embankment we may observe its outer slope down to and into the water and see that it is much steeper than the depositional slope of the top of the embankment. But this steeper slope lakeward is also a depositional slope, and we may safely infer that it has been so since the early stages of delta building. That the deposits on the flatter top slope rest on the edges of the more steeply inclined beds is also a safe inference. Looking lakeward from the outer edge of the embankment it is easy to infer that, out beyond the toe of the inclined beds, deposits of fine material are being laid down in layers parallel with the lake bottom. As the embankment grows lakeward the steeply inclined beds encroach on these flat-lying deposits and bury them.

Thus, from what I have said of the mode of deposit in a functional Lake Superior delta, anyone might have inferred the structure of deltas in general. But this is not the way we learned how deltas were built. It required a *post-mortem* examination to give us that knowledge. G. K. Gilbert about 60 years ago studied the basin of Lake Bonneville and described for us (1882) the shore features of the lake. The waters of Lake Bonneville at one time stood about 1000 feet above the level of the present Great Salt Lake. As the lake basin was emptied, down to the existing residual lake, the shore features of several of its stages were left high and dry on the mountain slopes of the desert. Among these were many deltas, and, as the surface of Lake Bonneville receded below the shore lines of which they formed a notable part, the streams cut trenches through them from top to bottom. This exposure of the whole delta structure enabled Gilbert to analyze and describe it in detail. He observed the threefold make-up of the embankments to which I have alluded in the Lake Superior deltas and he gave names to their constituent parts. He called the deposits which form the nearly flat surface the *top-set* beds, the more steeply inclined beds beneath these the *fore-set* beds, and the nearly flat beds below these the *bottom-set* beds. He also showed the relationship of these three structural elements in a diagram similar to that which is shown in Figure 1.

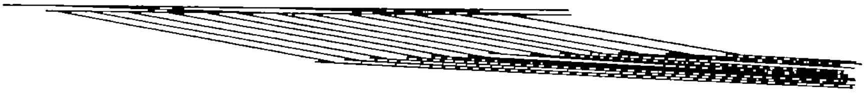

FIGURE 1.—*Diagram of structure of delta*

SURFACE SLOPES

This brief account of the small delta, based on Gilbert's classic work on the shore features of Lake Bonneville, serves equally well for the structure of the great deltas of the earth. The vast alluvial plain of the delta of the Mississippi, 386 km. in width, is continued seaward as a subaqueous slope for 129 km. to the 100-fathom contour, and the two together represent the expanse of the top-set beds. The subaerial portion of this expanse is nearly flat, and the subaqueous portion has a gradient of 1 in 710. Below the 100-fathom contour the outer slope of the delta has a gradient of 1 in 69 for another 195 km., down to the —3 km. (1640 fathom) contour. This is the region of deposition of the fore-set beds. For the next 180 km. seaward, down to contour —3.65 km. (2000 fathoms), the bottom-set beds have a gradient of 1 in 225. There can be little doubt but that, throughout the growth of the delta, the top-set beds have always been deposited across the upper edges of the fore-set beds at the front of the delta and that the latter have always extended out over the bottom-set beds.

THE ISOSTATIC PROBLEM

The most important feature of the Mississippi delta, from a geological point of view, is the great thickness of the embankment. It is recognized by geologists who know the delta best that the magnitude of its vertical dimension is due to the sinking of the old sea floor upon which the embankment rests. This subsidence has been a slow process and is probably still going on in the region of active growth of the embankment. It appears, moreover, to have been concomitant with the progress of growth without serious interruption from late Cretaceous time to the present. It is a short and easy step from the recognition of this association of sedimentation and subsidence to suppose that the sinking of the sea floor is due to the load of sediment which has been brought to it and dumped upon it by the great river—in other words, that the subsidence is isostatic. But isostasy means more than the depression of the earth's crust by a load imposed upon it. Under such load, whether it be due to continental glaciers, thrusting, or the growth of embankments, the sinking is an expression of disturbance of balance. To compensate for the newly imposed load heavy rock in the depth of the earth must flow away to other regions and so make depression possible; and the outflow not only creates depression, it maintains balance, or constancy of gravitative stress, in the column of the earth's crust beneath the disturbing load. For the region on which the delta is being built we may assume, for convenience in discussion, that at the beginning of the deltaic cycle it was in stable isostatic balance.

When we attempt to apply the isostatic mechanism to a sinking delta embankment, however, we encounter difficulties which inhere in the diversity between the density of the load imposed and that of the rock which flows away in depths. The latter has a specific gravity of about 3 3, while the former, even after compaction, does not exceed about 2.4. Thus a layer added to the delta, having a thickness of 3.3, will be balanced by the removal of a layer of heavy rock 2.4 thick. The outflowage of the heavy rock creates a depression which permits the bottom of the delta to sink 2.4, but its top rises by deposition $3.3 - 2.4 = .9$. It is obvious that the top of the delta is limited to the proximity of sea level and cannot rise higher by river deposit. The incompetence of the river to place its load above the level of its flood plain, at or close to sea level, appears to forever inhibit the embankment from depressing its floor beyond a very small initial amount.

But if we suppose a considerable depth of water to have been displaced by the delta, then the mechanism works. In general, if x be the thickness of the embankment in kilometers, when it comes to final balance, and d the initial depth of water, the sea level being supposed constant throughout the deltaic cycle,

$$\text{then } 2.4\,x - 1.026\,d = 3.3\,(x - d)$$
$$\text{whence } x = 2.52\dot{6}\,d$$

Any addition to the thickness in excess of $2.52\dot{6}\,d$ kilometers would cause an isostatic depression less than the added thickness so that the latter would pile up above sea level. The river cannot so pile its deposit, and $2.52\dot{6}\,d$ kilometers is the limiting value for the thickness of the delta.

The assumption of a constant sea level is, however, inadmissible. We must consider the effect of the change of sea level on the growth of the delta. The quantity of sediment derived from all continental erosion and of sea with deposition?

Initial [illegible] negative balance [illegible]

delivered to the sea during the time of growth of the Mississippi delta is so great that it has raised the level of the ocean by an amount that cannot be ignored. Elsewhere (Lawson, 1938) I have estimated that the rise of sea level due to this cause, within the time mentioned, is .6 km. This additional height of sea level means that the surface of the ocean is so much farther from the center of the earth than it was at the beginning of the deltaic cycle. During the rise the continental surfaces have suffered degradation. For every layer having a mean thickness of 3.3 removed a corresponding thickness of 2.67 has been added in depth, so that the distance from the continental surfaces to the center of the earth has been diminishing, while the sea surface has been rising. Should the two surfaces, thus approaching each other in position, ever coincide—*i.e.*, should extensive continental surfaces be reduced to peneplains—the rise of sea level would thereby be checked. The increase of distance of the sea surface from the center of the earth has come about slowly with the growth of the delta embankment and does not represent displaced sea water as does the initial depth of sea displaced in that growth. It does, however, determine an upward growth of the delta, also of .6 km. or 1968 feet. This greatly modifies the estimate just made for the limiting thickness of the delta.

If, as before, x be the thickness of the embankment at final isostatic balance, d the initial depth of the sea in which it was built, and k the rise of sea level in the deltaic cycle to the present time, then

$$2.4\,x - 1.026\,d = 3.3\,(x\text{-}d\text{-}k)$$
$$\text{whence } x = 2.52\dot{6}\,d + 3.6\,k$$

In the case under consideration $k = .6$ km. and $x = 2.52\dot{6}\,d + 2.2$.

PREVIOUS VIEWS ON SUBSIDENCE

In two excellent papers (Barton, *et al.*, 1933; Russell, 1940) dealing with the subsidence of the Mississippi delta this question of the initial depth of water is not discussed, although it is apparent that the authors consider the water to have been so shallow that they are unable to explain the great subsidence, which they make known, as a manifestation of the operation of isostasy. Barton and his associates state that since the beginning of the Tertiary there has been continuing progressive subsidence. Also they say (p. 1458):

> "But isostatically, the Gulf Coast geosyncline must be, and for a long time must have been, negatively out of equilibrium. Subsidence continued, however, and presumably must have increased the lack of isostatic equilibrium, as the progressive depression of the basement has increased the negative gravity anomaly. The movement, therefore, has been the reverse of what would be expected from the theory of isostasy. . . . The subsidence, therefore, seems more probably to be the effect

> of the sedimentation and to have tended to compensate it But the subsidence cannot be the effect of a movement toward isostatic equilibrium under the effect of extra load of sediments Superficial yielding of the basement under the weight of the extra load of the sediments seemingly must have taken place without regard to the isostatic relations."

Throughout the discussion from which these quotations are taken there is no reference to the depth of water displaced by the advancing embankment. The implication is that the authors considered it to be negligible, but they do not say even that much, although the water displaced is an important factor in the operation of the isostatic mechanism.

In Russell's (1940, p. 1227, 1228) fine paper, he says:

> "It seems difficult, if not impossible, to accept Louisiana evidence without coming to the final conclusion that the Gulf Coast Geosyncline is the result of sedimentary loading . . . The competence of sedimentary loading to depress material beneath and thus permit additional load at the original site seems clearly established by the lithologic and structural record"

But nowhere does he raise the question of the depth of water in which the delta embankment was built out until it attained mass enough to initiate depression. He seems, however, to be impressed with the fact that: "Continental, near shore, and shallow water sediments are typical of its whole Cenozoic column." Thus, although Russell recognizes the isostatic implications of depression under load, he, too, seems to regard the initial depth of water as negligible.

THE INITIAL LOAD OF SEA WATER

Like Barton and Russell I am impressed by the suggestion that, in the sinking of the Mississippi delta under its own load, we have an example of the operation of the principle of isostasy. I am also aware, as they seem to be without saying so, that, if the delta started in shallow water, we cannot invoke isostasy to explain the sinking. But, as far as I know, there is no reason whatever for the supposition that the delta was built out into shallow water. The counter assumption is that, when the delta started, there was a heavy load of sea water on the sea floor which was gradually displaced as the delta embankment extended out from the original shore. For the known thickness of the delta, 30,000+ feet south of New Orleans, the depth of such a sea could not have been less than 2.75 km. to provide the necessary mechanism for isostatic subsidence, which seems to exist in that part of the delta. Since, from the formula given above, if x be the required depth,

$$2.526\,x + 2.2 = 9.146 \text{ km. (30,000 feet)}$$
$$\text{whence } x = 2.75 \text{ km.}$$

But the 30,000 feet for the thickness of the delta to the south of New Orleans is a minimum value. I propose, therefore, to assume that the initial depth of sea in the northern part of the Gulf of Mexico was 4 km. and that the region was in isostatic equilibrium when that was the case. This estimate accords with the fact that soundings off coast lines aggregating 30,000 km. in Africa, India, Australia, North America, and South America show that in 61 profiles the —4 km. contour is distant from the shore on an average only 52 km.

There is nothing in the assumption which precludes the existence or, indeed, the prevalence, of shallow-water deposits throughout the entire geological column represented in the delta. Shallow-water sediments are a prime characteristic of all deltas, whether they are built out into deep water or shallow water; and nothing can be inferred from their presence in a depressed delta as to the depth of water antecedent to its displacement by the embankment.

For an assumed initial depth of water of 4 km., on a level sea floor, and a rise of sea level of .6 km. during the deltaic cycle, the limiting thickness of the delta, for all that extensive portion of it which is at, or slightly above, sea level, is $2.52\dot{6} \times 4 + 2.2 = 12.3$ km., provided, as already stipulated, that the earth's crust has in itself no supporting strength. But probably in no case would the thickness reach that limit. The load has to accumulate to a certain critical thickness, determined in large measure by the area over which it is spread, in order to inaugurate subsidence. This means that the more rigid crust, above the asthenosphere, has a certain very limited degree of supporting strength, inherent either in the arch of the earth's spherical structure, or in the frictional resistance to flow in the asthenosphere, when a layer of heavy rock is forced to evacuate its position, and so make subsidence possible. But, as the process of evacuation is excessively slow, the frictional resistance should be small. But, however small, it is probably the more important factor in giving to the crust its supporting strength in the early stages of loading.

THE SUPPORTING STRENGTH OF THE CRUST

So far as I am aware, we have no positive measure of the supporting strength of the earth's crust under load, apart from isostatic balance. The nearest approach to such a measure is Gilbert's (1890) suggestion that the emptying of Lake Bonneville caused an arching of the region by removal of load. In this case, accepting the suggestion, the Bonneville shore line, at an altitude of about .3 km. above the present Great Salt Lake, was arched when the water level dropped to the level of the Provo stage of the Lake, and then to that of Great Salt Lake. At the Provo stage, by reason of a relatively constant outlet, the water stood for a long time and developed magnificent shore features. But when the water in the basin evaporated down to the level of Great Salt Lake the abandoned Provo shore, about .18 km. above the present Lake, was also arched. The arching of the Provo shore was, however, not so acute as that of the Bonneville shore by removal of the load of water. We may infer, therefore, that a load of .18 km. of water is sufficient to depress the earth's crust in that region. In the absence of other data we may consider this load, if applied as broadly as it was in the Bonneville basin, as a limiting measure of the supporting strength of the crust. That is to say, any load equal to or greater than a prism of water .18 km. deep will depress the crust.

But even a large delta is a local affair, and the depression which accommodates it has traceable boundaries against the more stable region surrounding the embankment. It is difficult to avoid the conclusion that these boundaries, particularly those between it and the continental area, are zones of distributed shear, within which the subsidence is to some extent resisted by the strength of the crust.

Since we have no very definite measure of the competence of the crust, apart from isostasy, to support load we may to advantage compare two hypothetical cases. In the first the crust under the delta is supposed to have no supporting strength, or, to put it another way, the whole of the sedimentary load is effective for depression. Then, at final balance, when the embankment ceases to sink, if x be the total amount of depression,

$$2.4\,(x + 4.6) - 4 \times 1.026 = 3.3x.$$
$$\text{whence } x = 7.7.$$

The thickness of the embankment is then

$$7.7 + 4.6 = 12.3 \text{ km.}$$

In the second case a thickness y is assumed to accumulate on the sea floor before depression sets in. If x is the depression in kilometers, at final balance,

$$2.4\,(x + 4.6 - y) - 4 \times 1.026 = 3.3x$$
$$\text{whence } x = 7.7 - 2.\dot{6}y.$$

That is, for every 1 km. of load ineffective for depression the subsidence at final balance is $2.\dot{6}$ km. less than in the case where the whole sedimentary load is so effective.

If, now, we apply to the Gulf of Mexico the Bonneville measure of the supporting strength of the crust, and take the maximum value for the load necessary to depress the crust, the equivalent of .18 km. of water is .131 km. of delta sediment after compaction and correction for the levitation of sea water displaced. Then the depression at final balance is 7.7 — 2.6 × .131 = 7.35 km.; and the thickness of the delta is 7.35 + 4.6 = 11.95 km.—let us say, 12 km.

The only possibility of the realization of this hypothetical load, which may have been imposed on the sea floor without depressing it, is in the region of the slope of the fore-set beds, as it was in the early history of the delta. Today that slope is but little less than 200 km. wide, from north to south, and its upper or landward edge is over 500 km. south of the original shore line at which it started. In its very early stages this slope may have been somewhat steeper than at present, but it has always advanced seaward approximately parallel to itself, and, therefore, of course always risen by accretion parallel to itself. The sea floor immediately underneath this slope would, even in the very early stages of delta building, first feel the effect of loading in excess of support at a short distance seaward of the northern limits of the embankment. Subsidence once started would proceed steadily with accretion at the top in the form of fore-set followed by top-set beds; and the area of subsidence would widen with the seaward advance of the embankment. The limit of subsidence would have been reached when the sea floor, on which the delta rests, had been depressed 7.35 km. below the position at which it was when delta building started, reckoning positions with reference to the center of the earth. For at this depth the position of the bottom of the embankment would be stabilized by the impossibility of adding more load at the top.

SEAWARD EXTENSION OF THE DELTA

Thus the sinking of the site of the delta began at a very early stage of its growth, and, from that beginning to the final stage of stability, subsidence proceeded steadily with growth at the top. In this growth it is not clear to what extent the supporting strength of the basement, up to a limit of less than that represented by a load equivalent to .18 km. of water, continued to operate after that limit was overtaken by increasing load; but it probably represents a permanent stress difference in the isostatic mechanism. The stress difference being constant throughout subsidence, the latter should have been at a constant rate in any given vertical locus of sinking. Apart from this stress difference we may say that, at every stage of the subsidence, if we may speak of stages in a continuous process, the embankment has been in temporary isostatic balance. But it is pretty clear that the top of the delta could not permanently reach sea level in its upward growth until the bottom had reached a depth of 12 km. below sea level, at which depth it ceased to sink. This means that the shore line of the delta at any time is the locus out to which it has attained permanent balance. Both Barton and Russell appear to be of the opinion that subsidence is still going on. If this be so, it means, under the supposition of an initial depth of water of 4 km., that the embankment has not yet attained its limiting thickness of 12 km. But, in view of what has just been said, it seems improbable that this is so for the subaerial portion of the top of the delta. It seems more probable, since the sinking began at a very early stage of its growth, that the process involved the seaward extension of an embankment of the greatest possible thickness at all stages of that extension, out to the shore of the delta for every one of those stages. If this view be correct, it follows that the evidence of sinking cited by Barton, Russell, and others in the middle part of the subaerial delta plain is not a manifestation of depression now going on in that region, but is ancient history. The only part of the embankment now sinking, or capable of sinking, is probably the outer or seaward part, having a thickness of less than 12 km. All that part north of the shore has probably attained the maximum possible thickness and is stabilized in position until there comes an appreciable rise in sea level, which will make additions of new top-set beds possible.

THE BOTTOM OF THE DELTA

Here we may briefly consider the configuration of the bottom of the delta embankment. The profile of the top of the embankment is definitely determined by its relation to sea level. The land portion of the top, particularly, is the flood plain of a great river. Even if, in the mechanism of isostatic adjustment of disturbed balance, there were any tendency to arch the surface the tendency would be ineffective against the independent counteraction of the river and its distributaries; and any tendency to sag would be promptly countered by the filling of the depression by the river. Sagging has been, however, a very notable characteristic of the delta. As the embankment extends seaward, under an upward limitation at sea level for its top, the load becomes greater than the crust can sustain, and it becomes thicker by the sag of the floor on which it rests. The detailed evidence for this is set forth by Barton and by Russell in the papers heretofore cited. That evidence is unequivocal and satisfactory in the highest degree as to the subsidence of the embankment progressively with its growth. The increment of thickness comes to the embankment at its top, but accommodation for that increment is provided by depression of its bottom. As to the profile of the bottom Russell speaks of the general structure as being ladle-shaped, and the implication of his description is that the bottom is convex downward. Barton's figures show the embankment lying in a rather acute V-shaped trough, although the acuteness is largely due to the exaggeration of the vertical scale of the drawings. If the delta is in isostatic balance, and the initial depth of water was uniformly 4 km., then for the maximum possible thickness of the embankment—*viz.*, 12 km.—its bottom must be level. This condition could only have been arrived at slowly and progressively with the seaward growth of the embankment. The determining factor, at each and every stage of this outward growth, was isostatic balance at a thickness of 12 km. Beyond that thickness there was no tendency to further subsidence. The delta had then arrived at stability of position. This stabilization, as a dynamic process, would keep pace with the advance of the embankment seaward; and the bottom of the delta would be level out to a line parallel to and vertically below the present shore contour. From this line southward, the bottom of the delta would slope upward at a small angle. Beyond the line, southward, the embankment would be in unstable balance in the sense that it is still capable of taking further load, and so still further depressing the bottom upon which it rests. When the present fore-set slope, by continued accretion of sediments, becomes built up close to sea level and is buried by top-set beds at sea level, then the upward slope of the bottom of this portion of the delta will have sunk to a position 12 km. below sea level, and have become part of the level bottom of the embankment. For a rigid body this sinking would be effected by rotation on a pivotal line, at the southern extremity of the flat bottom of the older part of the delta, at a depth of 12 km. The angle of rotation would be that which the upsloping bottom of the now functioning fore-set portion of the embankment makes with the horizontal. This angle is 1°33′. But since the embankment is far from being rigid the sinking may possibly be effected by shear stress and corresponding strain, with small ruptures, distributed throughout the mass over which the load is becoming heavy enough to cause depression. Under the isostatic theory of deltaic subsidence this is about the only way in which a large embankment could advance seaward and maintain isostatic equilibrium throughout the process. Of course since a very early stage in the growth of the delta—*i.e.*, since the initiation of subsidence—the main embankment has had a seaward projection of less thickness, in the region of the fore-set and bottom-set beds, which gradually became incorporated in the main embankment as the latter advanced seaward. This incorporation was at all stages effected by addition of new sediments at the top and by the sinking of the whole enlarged mass by the rotation or distributed shearing above referred to. The final result, for the assumption of a uniformly deep sea to begin with, would be a level-topped, level-bottomed embankment in isostatic balance at the limiting thickness determined by the initial depth of water. At the front of this embankment, even in its final stage, would project those thinner underloaded portions, where deposition was still active, in the form of subaqueous top-set, fore-set, and bottom-set beds, and where movements toward final balance by rotation or shearing were still going on.

STRATIFICATION OF THE EMBANKMENT

I have called the top of the delta from the original shore line out to the 100-fathom contour, a width of 515 km., the region of the top-set beds. The subaerial portion of this has according to Russell (1940, p. 1216) an air-line slope of .32 feet to the mile. The strata in part of the top-set region have, therefore, a depositional dip of .32 feet per mile, and in the subaqueous part, 129 km. wide, the depositional dip is 600 feet in 129 km., or 80.15 miles—that is, 7.4 feet per mile. But while, under the hypothesis of an initial depth of water of 4 km. and a rise of sea level of .6 km., since the beginning of the deltaic cycle, the embankment has a thickness of 12 km., at the present shore line, and is in final isostatic balance, incapable of further subsidence, it may sink further between the shore and the 100-fathom contour. This subsidence will range from nothing at the present shore line to .18 km. at the 100-fathom line, when the shore shall have extended out to where there is now 100 fathoms of water. As the subaerial beds, dipping seaward .32 feet per mile, extend out over the present subaqueous beds, with their dip of 7.4 feet per mile, there will be an apparent disconformity between the two sets of beds. When the lower set has been thus buried, out to the present position of the 100-fathom contour, the added load will have caused the embankment to subside to its final position of equilibrium with a thickness of 12 km. In this settlement the block of the embankment between the present shore and the present 100-fathom contour will have effectively rotated through an angle the tangent of which is $\frac{.18}{129}$. Due to this rotation of the block the buried beds near sea level which had a depositional dip of 7.4 feet to the mile will have had that dip increased to about 15 feet per mile when they come to final rest.

If, however, by reason of certain constant characteristics, the whole of the deposit in this interval be considered a single stratum, the latter will have thickened steadily by accretion at the top, as well as by subsidence. When the shore line had reached the locus of the present 100-fathom contour its thickness there would be .362 km. (*i.e.* depth of water .18 km. and amount of subsidence .182 km.). If today a thin layer of volcanic ash were deposited on the sea bottom between the present shore and the 100-fathom contour, it would become a marker for the bottom of a wedge-shaped stratum. The dip measured on this marker would be .362 km. in 129 km., or 14.81 feet to the mile; and the top would of course be level.

On the fore-set slope, the depositional dip (gradient of slope) is $3 - .182 = 2.818$ km. in 195 km. or 76.3 feet per mile. But when, by seaward extension of the embankment, the 100-fathom contour moves out 195 km., and the fore-set beds are buried, a layer of ash, laid down now on the sea floor, would have a gradient or dip which is the sum of the present dip of the fore-set beds and their angular depression, that is, $76.3 + 135.6 = 211.9$ feet per mile. It is, however, improbable that these final relatively steep dips of the fore-set beds would ever be seen at the surface of the delta, although they might very well be encountered in bore holes penetrating the depth of the embankment. Except in cases of trenching, due to fall of sea level in glacial epochs, dips observable at the surface are those of the top-set beds, and the steeper of these would be marine strata, laid down off shore in shallow water and steepened by subsidence in the manner above indicated.

SNOUT OF THE EMBANKMENT

It has been pointed out that the embankment comes to balance when it is 12 km. thick, after allowing for a stress difference equivalent to a load of .131 km. of sediment. Following the seaward extension of the delta shore, the embankment progressively attains balance by continuous deposition of top-set beds in shallow water and corresponding depression. Beyond the shore the cross section of the embankment is determined by (1) the slope of the shallow-water bottom out to the brink of the fore-set slope, and the corresponding upward slope of its bottom to a locus immediately below that brink; (2) the relatively steep slope of the present sea floor where the fore-set beds are accumulating, and the corresponding upward slope of the old depressed sea floor below this; and (3) the flatter sloping sea floor in the region of bottom-set deposition, and the corresponding upward slope of the old depressed sea floor. These three elements of the profile of the embankment are all in temporary balance for the masses which they respectively contain. All three are, however, susceptible to further subsidence as depositional load is added to them, to the limit of 12 km. below present sea level, continuing the same horizontal bottom and uniformity of thickness as prevail to the landward of the existing delta shore. The configuration of the profile of the delta at all stages, after final balance has once been attained in its early growth, is that of a flat-bottomed frying pan, rather than a ladle. The profile is, however, asymmetric. The steep side is to the north, seaward of the original shore line. The less steeply inclined side of the pan is in three straight segments, nearly symmetrical in gradient to the subaqueous top-set, the fore-set, and the bottom-set slopes, respectively. In general the profile beyond the shore is angular both at top and bottom, tapering seaward. It may be called the snout of the delta. The length of the snout of the delta is greater than the main body of the delta now in permanent balance, but its volume is less. The delta grows by the gradual conversion of the snout into a prism uniformly 12 km. thick.

SLOPING SEA FLOOR

The discussion of the bottom of the delta has been limited thus far to conditions flowing from the supposition of uniform initial depth of water—*i.e.*, a level sea floor in the region now covered by the delta. But it is, of course, entirely possible that the sea floor may have had a down slope seaward. Let us suppose that a portion of the sea floor had a gradient of 1 per cent. At the upper end of this slope the initial depth of water may be taken at 4 km. Then at a distance of 100 km. seaward the depth would be 5 km. As the embankment advanced seaward the slope would be buried by sediment about as fast as the level sea bottom, but it would take longer for the top to arrive at sea level even if there were uniform subsidence. But when the thickness of embankment on the lower part of the slope became greater than on the more shallow part, as it would in the course of time, the increasing load would doubtless accelerate the rate of subsidence and retard the time for completion at the top. In this case, if the curve of acceleration were not rectilinear, increase of load might cause a curvature of the buried sea floor—*i.e.*, of the bottom of the embankment. But whether that bottom were curved or straight, the thickness of the out flowage of heavy rock in depth for compensation would be greater beneath the lower part of the slope than beneath the shallower part. Its upper boundary in profile at final balance would be symmetrical with the bottom of the embankment—*i.e.*, whereas the latter at final balance would slope down seaward, the former would slope upward.

DIMENSIONS OF THE DELTA

The boundaries of the delta cannot, of course, be stated with exactitude. In an attempt to estimate its approximate area the northern boundary, though rather sinuous, is taken to be at Lat. 32°30′. The southern limit is a fading boundary on the sea floor north of Sigsbee Deep, or perhaps in that depression, where the bottom is very flat. The 2000-fathom contour, on the north side of the deep, may be taken as a convenient and conservative limiting line. In Long. 91° W. this line coincides with Lat. 24°18′ N. The delta thus covers over 8° of latitude, its width being 879 km.

The land surface of the delta from its northern boundary to the shore, at about Lat. 29° N., measures 386 km. on a N.-S. line. From the shore to the 100-fathom contour, at Lat. 27°57′ N., the distance is 129 km. These two portions of the delta, totaling 515 km. in width, comprise the flat-lying top-set beds of the embankment. The slope from the 100-fathom contour to the 1640-fathom (—3 km.) contour, comprising the present surface of the fore-set beds, is 195 km. in width and has a gradient of 1 in 70. From the 1640-fathom to the 2000-fathom contour the width is 169 km., and the gradient is 1 in 256. This slope is in the region of the bottom-set beds, which fade insensibly into the deep-sea deposits.

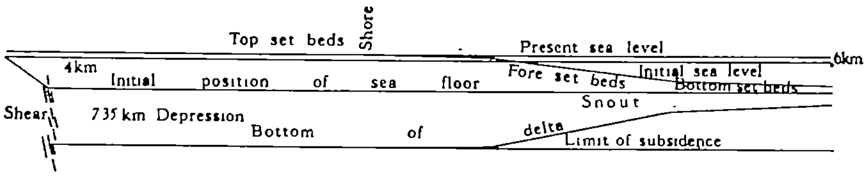

FIGURE 2.—*Diagrammatic north-south cross section of the Mississippi delta*

Showing initial and present positions of sea level, initial position of sea floor at an assumed depth of 4 kms., and the present bottom of the delta after depression to the limit of 7.35 kms At the landward side of the depression a shear is suggested. Seaward of the shore, the delta, not yet depressed to the limit, tapers in the form of a snout

Since the initial depth of water limits the thickness, and the depth is assumed to have been uniform over the northern part of the Gulf of Mexico, the bottom of the embankment is as level as the original sea floor. We are thus in a position to construct a hypothetical profile of the embankment, and such a profile is shown in Figure 2. The cross-sectional area of the profile is 8260 square kilometers. If this profile, which is drawn about on Long. 91° W., be taken as characteristic of the embankment throughout its length, we may arrive at approximate figure for the volume of the delta. The embankment extends from about Long. 87° W. to Long. 99° W., a distance of 830 statute miles or 1336 km. Russell speaks of "the geological processes which have been dominant since the late Mesozoic in forming a sedimentary section some 30,000 feet thick, extending coastwise for over a thousand miles." For the more limited figure which I have just given for the east-west extent of the embankment its volume is $8260 \times 1336 = 11{,}035{,}360$ cu. km.

SUPPLY OF DELTA SEDIMENTS

This large quantity of sediment has been brought to its present position by the agency chiefly of the Mississippi River. It is derived from the hydrographic basin of the river and represents about half the total erosional product. The work of Dole and Stabler (1909) supplement-ing the earlier work of Humphreys and Abbott (1861) has shown that the surface of the basin of the Mississippi River is being denuded at the rate of .056 mm. per year.[1] Of this amount Dole and Stabler have shown that 25 per cent goes to the sea in solution and 75 per cent as solids. The present writer estimates that one-third of the solids goes to the ocean shores and bottom beyond the limits of the delta, and that two-thirds, or one-half of the total erosional product, goes to the construction of the delta.[2] Thus the total volume of the delta after correction for difference of density is somewhat greater than one half the volume removed from the area of the river basin. The erosional removal is

$$11,035,360 \times \frac{2.4}{2.67} \times 2 = 19,838,848 \text{ cu. km.}$$

Since the area of the basin is 3,276,274 sq. km., the thickness of the layer removed is $\frac{19,838,848}{3,276,274} = 6.03$ km., let us say 6 km.

INITIAL MEAN ALTITUDE OF MISSISSIPPI BASIN

For removal of a layer 6 km. thick from the area of the basin and the delivery of the same to the ocean, the isostatic rise of the eroded surface is $\frac{2.67}{3.3} \times 6 = 4.854$ km. The lowering of the surface of the basin is $6 - 4.854 = 1.146$ km. On the basis of a series of hypsometric sections drawn across the basin I have estimated that its mean altitude above present sea level is .80 km., or $.8 + .6 = 1.4$ km., above the position of sea level at the beginning of the deltaic cycle. From these data we may ascertain the mean altitude of the surface of the basin at the beginning of the cycle. Since the present mean altitude is .8 km. and the lowering of the surface is 1.146 km., the initial mean altitude above present sea level is $.8 + 1.146 = 1.946$ km., or $1.946 + .6 = 2.546$ km. above sea level as it was at the beginning of the deltaic cycle.

FLUCTUATION OF SEA LEVEL

If the region of the Mississippi delta were to be slowly uplifted 300 or 400 feet what would be the effect on its land surface? First of all the various rivers flowing to the sea across it, trunk stream and distributaries, would have their grades steepened and their velocity increased. As a consequence their corrasive power would be greatly enhanced. Every stream would tend to cut a trench for itself and, if the uplift were not too slow, would do so. No distributary could branch out from such a trench. The depth of the trench would be closely determined by the change in the relative position of sea surface and delta surface. The antecedent surface of the delta would become a broad terrace traversed by the trenches. In time, when the streams had re-established their proper gradients in these trenches, they would proceed to widen them by lateral corrasion and evolve broad bottoms, thus accentuating the terraced aspect of the topography. Holding this picture in mind let us consider now what would happen if the region slowly subsided, till the former hypsometric relation of sea and land were re-established. How would the physiography of the delta surface change? The gradients of the streams in all these broad terraces would be lowered, first at their mouths by building little subdeltas, and thence gradually upstream. Eventually the trenches, no matter how broad, would be filled to the brim. Then every stream would build up levees a little higher than the plain on either side, and would sooner or later spill over to form new distributaries. The surface of the delta would have been restored to its normal character.

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[1] For summary of data, see Lawson (1934)

[2] There is, however, very little observational basis for this estimate.

In the later geological history of the Mississippi delta, according to the brilliant work of R. J. Russell, this trenching of its surface to depths of from 100 to over 300 feet, followed by filling of the trenches, has occurred no less than five times. Frink (1941) confirms Russell's earlier statement regarding this five-fold recurrence of degradation and aggradation of the delta surface. Accepting Russell's reading of this deltaic history we might conclude from what I have said above that the region of the delta had been subjected to a five-fold recurrence of uplift and depression, thus recognizing an extraordinary mobility of the earth's crust. But our conclusion would be fallacious. The phenomena of the recurrence of alternate degradation and aggradation may equally well be explained by the rise and fall of sea level. Students of continental glaciation are well aware that the glacial epoch was characterized by a sequence of glaciations and deglaciations. At every glaciation water was withdrawn from the ocean and at every deglaciation that same water, or a fraction of it, was restored to the ocean. There was thus a sequence of rises and falls of the sea surface in the latter part of the time in which the Mississippi delta was built. The failure, until recent years, to recognize the magnitude of this alternate rise and fall of sea level in Pleistocene time is an interesting phenomenon of the progress of geological thought. One which should make us blush. A great many geologists, including myself, have seen San Francisco Bay and the Golden Gate. Up till quite recently scarcely a man of them but heaved a great sigh of satisfaction and exclaimed: What a splendid proof of the depression of the coast! And yet, although the bay is certainly flooded from the ocean by way of the Golden Gate, the inference that a depression of the coast caused the flood is not necessary. The rise of the level of the ocean would do the job quite as well. Today it is impossible to disbelieve that in Recent time the surface of the whole ocean has risen a very notable amount. Two years ago I made a somewhat elaborate computation (1940) of the rise of sea level due to the melting of part of the ice of the last continental glaciation. I came out of the daze with a figure of 426 feet for the measure of the rise. The estimate, of course, lacks precision, and I do not regard it as more than a first approximation. But however it may be discounted there is no escape from the conclusion that the melting of the ice raised the level of the ocean several hundred feet. The deepest sounding in the Golden Gate is 414 feet, and the position of that sounding is probably the bottom of a gorge close to the old shore line. The sea level, when the continental glaciation was at its maximum, must have stood 414 + feet below the present sea level, which is fairly close to 426 feet.

Returning now to the Mississippi delta, the five-fold recurrence of degradation and aggradation of its surface is correlated by Russell with a corresponding recurrence of fall and rise of sea level, and the cause of the latter is ascribed to glaciation and deglaciation. Other workers in the same field, particularly Barton, Frink, and Fisk, are in agreement with him. The stratigraphy is arrived at by the inspection and measurement of formations in natural exposures in the upstream extension of the delta, and following these seaward by means of the record of wells. There are over 100,000 wells, many of them very deep, drilled into or through the delta. The record of these, though not always satisfactory, together with core samples, have in most cases been preserved. A study of these shows that the formations all dip seaward and increase in thickness and in angle of dip in the direction of the dip. The formations referred to are, for the most part, the deposits that were contributed to the delta in times of high sea level, when the process of sedimentation was exceptionally active in what is now the central part of the delta. The increase in thickness of several of these formations, as they are followed down the dip, is very remarkable. Three of them thicken to over 3000 feet each, a fact which impressively proves gradual subsidence *pari passu* with deposition.

Owing to the inconstancy of the stream courses across the surface of the delta—that is, their habits of migrating from one position to another in the course of long periods—the center of maximum deposition also shifts. It results from this that the structure of the delta is not simple, but is made up of an overlapping aggregate of many deltas, each one of which has its own center of accumulation for the time being, and, of course, also its own center of subsidence. But that is a detail upon which I am not qualified to speak. I am concerned here with the embankment as a whole.

Meanders of a delta [illegible]
may also be done by [illegible]
deglaciation [illegible]

But the migration of the streams on the delta surface has much to do with the formation of terraces, and the preservation of these from burial in areas of minimum deposition, as well as in times of high sea level, after the streams have become adjusted to that level. Russell states that the oldest of these terraces slopes seaward at about 10 feet to the mile; the next younger slopes 6 feet, the next 2 feet, and the next 1 foot to the mile. The most recently formed terrace, corresponding to the present high sea level, slopes only half a foot to the mile. These slopes, however, steepen toward the central part of the delta. The slope of the terraces, in excess of about half a foot to the mile, and the increase of that excess with age, is taken to be an expression of the subsidence of the main embankment. But, from what I have already said, such subsidence probably occurs to the seaward of the delta shore, the landward side of that line having arrived at stability of position, except for sinking due to compaction.

The vicissitudes which mark the later stages of growth of the delta of the Mississippi are, as Russell states, referable, with little or no doubt, to the recurrence of glaciation and deglaciation, and the effect of these on sea level, in Pleistocene time. In Tertiary time we have no evidence of glaciation in arctic regions. On the contrary fossil plants in certain parts of the arctic indicate a warm climate. So we may picture to ourselves a preglacial condition of the earth when arctic and antarctic regions were devoid of ice. Then the ocean basins were full to the brim. In the time of maximum Pleistocene glaciation enough water was taken from the ocean to lower its level probably 800 feet below the preglacial position. One effect of this drop of sea level was the emergence of the Mississippi delta and of all other marine deltas outside of the frozen terranes. The surfaces of all these deltas, we may be sure, were deeply scored by stream trenches. This, however, was a temporary condition. The next interglacial epoch was surely recorded in the delta by the converse set of conditions. This alternation of glacial and interglacial climate was recorded in changes of the position of sea level which affected all the coasts of the earth. Thus in the subtropical region of the Gulf of Mexico the relative chronology of the major events of the glacial period are being deciphered far more clearly and definitely than they have ever been in subarctic latitudes.

END OF THE DELTAIC CYCLE

As to the close of the deltaic cycle of the Mississippi there is this to be said: Geologists believe that the most characteristic structural feature of many mountains is the folded condition of sedimentary strata. The folding is in all cases due to horizontal compression, but there are all degrees of intensity of folding, from broad, shallow undulations to highly compressed, intricately contorted and overturned plications. They have observed also that the thickness of sedimentary strata in the mountains is much greater than elsewhere, and that the dominant feature of many folded ranges is a great syncline, cradled in which are many subordinate synclines and anticlines. The dominant syncline outlines a mountain range and is known as a geosyncline. Such a structure is well exemplified in the Appalachian Mountains, where it has been shown by stratigraphic studies that the thickness of shallow-water beds is much greater in the axial region of the vast troughlike mountain fold than it is on its flanks. It is evident that, in the depositional field, where this body of strata accumulated, the earth's crust subsided to accommodate the deposits. The maximum subsidence, and therefore the maximum volume of sediments, was localized along a zone which became, by mountain-making deformation, the axis of the geosyncline. There can be but little doubt that the subsidence was due to yielding underload, and that, throughout the sinking, crustal balance was maintained by out flowage of heavy rock in depth from beneath the seat of deposition. The maximum thickness of sedimentary strata in the Appalachian Mountains is about 40,000 feet. The attainment of this volume seems to have brought the sedimentary process to an end. The slow subsidence of accommodation to maintain isostatic balance gave place to a more vigorous deformation due to the collapse of the depositional trough under horizontal compressive stress. A mountain-making movement was thus inaugurated. The trough became a geosyncline, and this by continued compression was hoisted high above sea level and sunk deep into the crust. The range thus created was supported by the depression of the keel of the geosyncline. Heavy rock in depth was displaced, and the new range, consisting of light rock, acquired the buoyancy necessary for its existence as a feature of the surface relief. But the range was hoisted into the zone of vigorous erosion and in the course of time became greatly reduced. Its reduction signified a loss and surface redistribution of its mass. The buoyancy of the range then operated for its uplift, so that the lowering of its surface was far less than the mean thickness of the prism removed by erosion. The uplift, of course, also applied to the bottom of the geosyncline, so that the keel is not nearly as deep as it was originally. The key to this sequence of events in the genesis of a mountain range like the Appalachians appears to be that, as accumulation and subsidence proceed, several miles of the upper part of what seismologists call the "granitic" layer are displaced by weak incoherent sediments. The lower part of the "granitic" layer as it sinks approaches the asthenosphere and loses strength as it does so.

The residual upper part may be much less than half the thickness of the normal "granitic" layer, and be so attenuated that it yields to the compressive stress inherent in the crust. The maximum thickness of strata in a single sequence known anywhere appears to be 40,000 to 42,000 feet; and we are justified in the inference that when this thickness is attained further additions are precluded by the inauguration of the mountain-making movement.

The foregoing outline of the genesis of the Appalachian range, and of the antecedent sedimentary cycle, is indicative of the fate that awaits the Mississippi delta. The next mountain range to be developed at the continental margin will have an east-west trend, between south Texas and north Florida. The Mississippi River will have to find another outlet for the drainage of its basin, perhaps between the east end of the new range and the south end of the Appalachians. The present deltaic cycle will have closed. How long it may be in terms of years before the end it is hard to surmise. If the Mississippi delta started in a sea 4 km. deep, it now has a thickness very little less than 40,000 feet. Under that assumption it is evidently approaching the critical stage. But under the same assumption it can never get any thicker, and may be able to withstand the comprehensive stress of the crust until it is much broader than it is now.

SUPPLEMENT

As a supplement to the conclusions arrived at by this study of the Mississippi delta I may add another of general interest. On two oceanographic charts of the earth the depth of the ocean is indicated by submarine contours. The highest of these, at a depth of 200 meters or 656 feet, encloses the continental and insular areas. It includes the extent of off-shore water having less than that depth. The shallow seas thus indicated are particularly noteworthy off the coasts of eastern and southeastern Asia and the East Indies, Australia, Antarctica, eastern North America including the Gulf of Mexico, western India, as well as Bering Sea and the adjoining Arctic Ocean, Hudson Bay, and the Persian Gulf. I have told you that the melting of the ice of the Glacial Period has thus far raised the level of the ocean over 400 feet. But only about half of that ice has been melted. The other half still remains. If it too should melt, by further amelioration of world climate, the surface of the ocean would rise a few hundred feet more. We may safely infer, therefore, that in preglacial time, when there was no ice, the coasts of all the oceans must have been inundated more extensively than they are today. This inundation would of course have determined a configuration of coast line very different from that with which we are now familiar.

The difference would have been most marked in coastal regions of low relief with large river valleys; while bold coasts presenting mountain fronts, like that of the west side of South America, would not have differed much from those of the present time. We may also safely infer that, at the time of maximum glaciation, the greater part of the sea bottom within the — 200 meter contour would have been exposed as dry land, thus adding very considerably to the area of the continents. This addition of flat land to the continental margins would have induced migration of people then living on the earth to the new land surface. They would have moved in slowly as the shore retreated and probably would not have been conscious of the change going on. There they flourished for many centuries, if not millennia, with all the material adjuncts of a primitive social order. By the rise of ocean level, that came with the melting of the continental ice sheets, that migration was reversed. The people moved up the coastal slope in front of the rising tide, also slowly and also probably unconscious of the change that was going on. The material evidences of their existence are now lost to us. Their habitations, their monuments and their graves, together with their art and their refuse, lie scattered over the floor of the ocean. All that is left to us of this flight from the sea is the myth of a universal flood which we have never believed.

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PALEOZOIC PALEOGEOGRAPHY OF ARIZONA

BY ALEXANDER STOYANOW

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ABSTRACT

In Arizona two major areas of deposition separated by Mazatzal Land had already been established by late pre-Cambrian time. The first Paleozoic transgression from the northwest, starting possibly late in the Early Cambrian and depositing the Tapeats sandstone, reached central Arizona The southern limit of the Bright Angel shale and the overlap of the Muav limestone on the Tapeats indicate subsequent movements northward and southward in early Middle Cambrian time before the final withdrawal of the Cambrian sea from northern Arizona. In southeastern Arizona, likewise, only the earliest phase of the Cambrian transgression reached the central part of Mazatzal Land; the northwestern boundaries of the late Middle Cambrian Troy, Santa Catalina, Cochise, and the Upper Cambrian Abrigo formation are all successively regressive southeastward. Ordovician strata are known in Arizona only in the areas near New Mexico Silurian and early Devonian deposits are present only near the extreme peripheral margins of Mazatzal Land beyond the limits of the State. Central-western Arizona was a positive element during the Ordovician, Silurian, and both the Early and Middle Devonian. The overlapping late Devonian transgression, also initiated from the northwest and southeast, deposited clastics around the central part of Mazatzal Land and was succeeded by deeper sea conditions in Mississippian time. The southeastern trough was the deeper of the two basins throughout the Late Paleozoic and is regarded as an arm of the Ouachita seaway. Following the retreat of seas from most of Arizona in the Late Mississippian a marked early Pennsylvanian transgression was introduced from the southeast and limited on the northwest by Mazatzal Land. Grading of the Manzano gypsiferous series and fossiliferous limestones into the Supai northward and distribution of Kaibab faunas establish the relation of the southeastern trough to the northern area of Permian shallow marine and continental deposition.

INTRODUCTION AND ACKNOWLEDGMENTS

The writer (Stoyanow, 1936) has briefly discussed the Paleozoic waterways of Arizona; to outline these seas required many additional observations. Dr. G. M. Butler, former Director of the Arizona Bureau of Mines, enabled the writer to study, in connection with the summer programs of geological work of the Bureau, some of the most critical areas. Grateful acknowledgments are also due Dr. Eldred W. Wilson, Geologist of the Bureau, who willingly helped both in the field and in the office, and Mr. L. F. Brady, who assisted the writer on many of his trips. This paper discusses the distribution of Paleozoic basins and lands in Arizona on the basement complex and their relation to the main neighboring seaways and land masses. The accompanying maps are based on factual outcrops and distribution of strata. The imperfections of many of the postulated views are fully recognized. It is hoped that the facts presented will contribute toward the interpretation of some of the major paleogeographic problems of the Southwest, and a critical evaluation of suggested solutions will be greatly appreciated.

BASEMENT COMPLEX OF ARIZONA

Ransome (1916, p. 133) separated Arizona topographically into three principal regions: the plateau region which occupies the entire northern and northeastern parts of the State; the mountain region which extends from southeastern Arizona northwestward beyond the central part of the State, roughly paralleling the southwestern margin of the plateau region; and the desert region of southwestern Arizona. For paleogeographic studies, Arizona is readily separated into two regions: (1) the plateau on the northeast, where the thick, nearly horizontal Paleozoic and Mesozoic strata cover the basement complex except in the Grand Canyon area, and (2) the vast platform made up of schist and granite which composes the southwestern half of the State (Fig. 1). The trend of the main Paleozoic waterways on this platform was toward central Arizona from the northwest and, approximately along the mountain belt of Ransome, from the southeast. Throughout the southwestern region stratigraphic contacts and interrelation of rocks can be studied with minutest detail, whereas the paleogeography of northeastern Arizona, with much evidence buried beneath the plateau, can only be surmised from the studies of the rocks along its southwestern rim and in the Grand Canyon.

L. F. Noble (1914) studied the Archeozoic basement complex and Algonkian strata of the Grand Canyon area and, in the words of Ransome (Noble, 1914, p. 8),

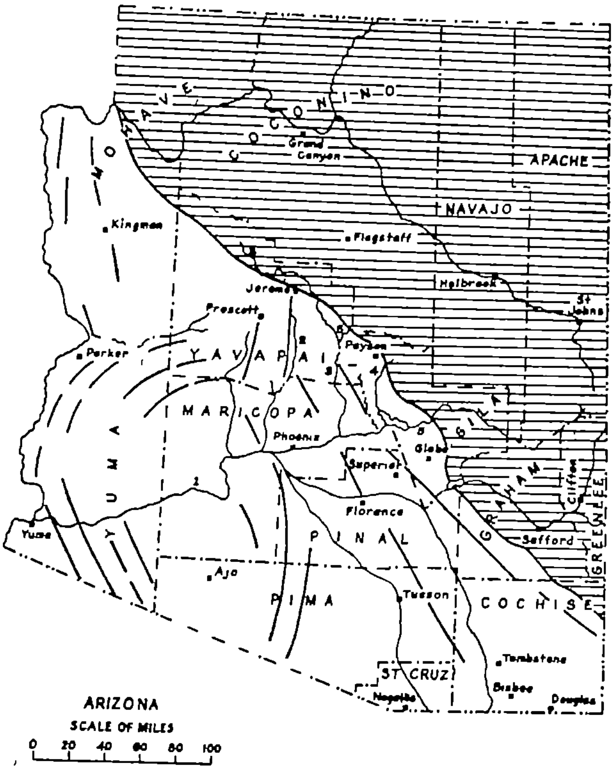

FIGURE 1.—*Orientation map of Arizona*

Basement platform unshaded, trend and virgation of the schistose and granitic mountains indicated by black lines. Plateau region with nearly horizontal strata ruled. (1) Gila River, (2) Agua Fria, (3) Verde River, (4) Tonto Creek, (5) Salt River, (6) Natural Bridge.

> "has not only thrown light on the pre-Cambrian history of the Colorado Plateau region, but has supplied geologists who are working in the southwestern part of the country with a standard of comparison for Algonkian strata exposed elsewhere in that region."

The major facts that seem to be firmly established in the pre-Cambrian geology of the Grand Canyon are that (1) only the Archean rocks were affected by the great granitic intrusion, (2) a profound erosion interval separated the Archeozoic rocks from the Algonkian, and (3) a profound erosion interval with structural adjustments elapsed before the introduction of Cambrian waters.

Two contributions have been made recently regarding pre-Cambrian rocks and geological events of central Arizona, paleogeographically the most important part of the State.

N. E. A. Hinds (1935, p. 27-29; 1936, p. 92-102; 1937, p. 1994-1998) outlined the geological development in central Arizona during the pre-Cambrian which he based partly on previous, and partly on his own, observations. As in the Grand Canyon, the Archeozoic rocks had been tilted, intruded by granite, and eroded before Algonkian time. In Algonkian time eruptions of rhyolite preceded the deposition of the Mazatzal quartzite, one of the principal pre-Cambrian members of central Arizona; orogeny, another granitic intrusion, and erosion followed.

Wilson (1939), who studied the central Arizona area in detail, offered a different interpretation. Volcanics, Yaeger greenstone, and Red Rock rhyolite are at the base of the pre-Cambrian complex. On the rhyolite are partly metamorphosed shale and grit of the Alder series which Wilson (1939, p. 1122) correlated with the sedimentary portion of the Yavapai schist of central Arizona. The Yavapai schist is usually placed with the Vishnu schist of the Grand Canyon and the Pinal schist of southeastern Arizona in the pre-Cambrian (Jaggar and Palache, 1905; *compare* Wilmarth, 1938, p. 2383), older pre-Cambrian (Ransome, 1916), or pre-Cambrian crystalline rocks (Darton, 1925). Above the Alder series more quartzite and shale is followed by the Mazatzal quartzite which, however, in the headwaters of the East Verde River and of the Tonto Creek rests directly on the Red Rock rhyolite. Wilson found evidence for only one granitic intrusion in central Arizona—of post-Mazatzal quartzite time; he gathered ample proof of the effect of a granitic intrusion on both the Red Rock rhyolite and the Mazatzal quartzite.

Wilson's evidence is opposed to the development of the pre-Cambrian complex in other regions of Arizona. The difficulty in his interpretation is that, like the Laurentian granite, the Yaeger greenstone and Red Rock rhyolite do not rest on any known foundation, but unlike that classic granite they are so limited areally as to cause doubt about their role, which they may be construed to share with the Archeozoic schists, of the most primordial rocks of Arizona; they have no counterparts in the Grand Canyon or elsewhere in the State.

The studies of geological sections in various parts of the Grand Canyon have proved the existence, within Arizona, of pre-Algonkian granite. The Algonkian rocks there, between the Archeozoic schist and Cambrian sedimentaries, have not been affected by the granite which intruded the schist.

Wilson's data accumulated in central Arizona admit of three theoretically possible solutions: (1) The Mazatzal quartzite, together with the older greenstones, rhyolites, and the equivalent of the Yavapai schist, all intruded by the same granite, is older than the Archeozoic Vishnu and Pinal schists and is one of the oldest members of the basement complex of Arizona, in which case Wilson's contention of only one major pre-Cambrian granitic intrusion is sustained (Table 1). Against the Archeozoic age of the Mazatzal quartzite, however, is the fact that in many places the Mazatzal is in no way more altered than the similar sediments of the Grand Canyon series, for instance the normal conglomerates within it in the East Verde headwaters. (2) The Mazatzal quartzite is younger than the Archeozoic schist but older than the strata of the Grand Canyon series. Assuming again only one major granitic intrusion it follows that the development of the Mazatzal quartzite with preceding volcanic activities is a chapter of the pre-Cambrian history not represented in the Grand Canyon area or elsewhere in Arizona. Such a postulation would call for a widespread removal of these or equivalent rocks, all intruded by granite simultaneously with the Archeozoic sedimentaries, with the evidence of their existence preserved only in a limited area of central Arizona. This alternative, however, admits either of one (post-Mazatzal) or, as proposed by Hinds, of two (post-Vishnu and post-Mazatzal) granitic intrusions and erosion intervals (Table 2). (3) The Mazatzal rocks—*i.e.*, volcanics, schist, quartzite, and intrusive granite—

TABLE 1.—*Pre-Cambrian of Arizona. Interpretation* 1

Southeastern Arizona		Central Arizona			Northern Arizona
			Devonian		
Cambrian	Cambrian	Cambrian			Cambrian
Unconformity					Unconformity
Apache group				Apache group	Grand Canyon series
Ep-Archean granitic intrusion and erosion interval					
Pinal schist					Vishnu schist
		Mazatzal group			

are of the same age as, or younger than, the Grand Canyon series but represent an extremely localized development confined to central Arizona (Table 3). The Mazatzal quartzite overlaps the older sedimentary rocks and in the Payson area of central Arizona rests with a basal conglomerate directly on the Red Rock rhyolite (Wilson, 1939, p. 1125); this may indicate a previous erosion with subsequent deposition of sediments and orogenic processes, culminating in a second granitic intrusion similar to other local and later granitic intrusions observed elsewhere in Arizona (Wilson, 1937, p. 1999).

TABLE 2.—*Pre-Cambrian of Arizona. Interpretation 2*

Southeastern Arizona		Central Arizona			Northern Arizona
Cambrian	Cambrian	Cambrian	Devonian		Cambrian
Unconformity					Unconformity
Apache group				Apache group	Grand Canyon series
Post-Mazatzal granitic intrusion and erosion interval					
		Mazatzal group			
Post-Vishnu granitic intrusion and erosion interval ?					
Pinal schist					Vishnu schist

The important fact in this connection is that, both in the Payson area of central Arizona at the headwaters of the East Verde River and at the Tonto Creek, and in the area north of Jerome in the basin of the main Verde River, the Cambrian Tapeats sandstone rests on the remarkably smooth surface of large masses of granite or on the uniformly tilted schist (Pl. 1, fig. 1). On the other hand, its contacts with the Mazatzal quartzite (Wilson, 1922, Pl. 31) are very much like the contacts of later Devonian strata with the same quartzite—the younger strata abut against or overlap the quartzite tilted at various angles within short distances. These two different relations of the Tapeats sandstone with the granite and the quartzite are strikingly similar to its relations with the smoothed surface of the Archeozoic and the rugged surface of the Algonkian of the Grand Canyon area which have been recently described by Sharp (1940).

The writer is of opinion that the extensive granite, closely associated with schist, that forms the basement complex of the entire southwest of Arizona is pre-Algonkian.

TABLE 3—*Pre-Cambrian of Arizona. Interpretation* 3

<table>
<tr><th colspan="2">Southeastern Arizona</th><th colspan="3">Central Arizona</th><th>Northern Arizona</th></tr>
<tr><td>Cambrian</td><td rowspan="5">Cambrian</td><td rowspan="3">Cambrian</td><td rowspan="3">Devonian</td><td rowspan="2"></td><td>Cambrian</td></tr>
<tr><td>Unconformity</td><td>Unconformity</td></tr>
<tr><td rowspan="3">Apache group</td><td>Apache group</td><td rowspan="3">Grand Canyon series</td></tr>
<tr><td colspan="3">Post-Mazatzal granitic intrusion and erosion interval</td></tr>
<tr><td colspan="3">Mazatzal group</td></tr>
<tr><td colspan="6">Ep-Archean granitic intrusion and erosion interval</td></tr>
<tr><td colspan="2">Pinal schist</td><td colspan="3">Yavapai schist ?</td><td>Vishnu schist</td></tr>
</table>

Ransome (1916, p. 165-166) believed that a natural barrier, probably a submarine ridge, existed in central Arizona in Cambrian time and possibly throughout the Paleozoic; this concept has been gradually developed, with further research in stratigraphy and paleogeography, into the idea of a land mass—Mazatzal Land, a positive element that separated the two principal areas of deposition in Arizona all through the Paleozoic (Stoyanow, 1936, p. 462) and also had a definite bearing on the Mesozoic paleogeography of the State.

MARINE INVASIONS OF CAMBRIAN TIME

As early as late pre-Cambrian time the southeastern and northwestern areas of deposition seem to have been well established in Arizona. In Cambrian time, Mazatzal Land was encroached upon first from the north by the shallow Tapeats sea, the deposits of which are found farthest south of all three members of the Grand Canyon Cambrian, or the Tonto group, and are seen at present in the area of the East Verde headwaters in central Arizona (Pl. 5a). The age of the Tapeats sandstone, formerly interpreted as very early Middle Cambrian (Stoyanow, 1936, p. 480), may be partly or entirely Early Cambrian, depending on the relation of the shales with *Olenellus* found in the western part of the Grand Canyon (McKee, 1937, p. 341) to the main body of the sandstone.

In *Anoria-Glossopleura-Alokistocare*, or earlier Middle Cambrian, time the northwestern sea retreated, and the corresponding formation, the Bright Angel shale, is not found very far south of the Grand Canyon. This retreat was followed by another transgression southward. The Muav limestone overlaps the Bright Angel shale, and at its southernmost observed termination, which is near Fort Rock, half way between the Grand Wash Cliffs and Prescott, it rests directly on the Tapeats sandstone.

The Cambrian waters began to invade southeastern Arizona only after these events. The Troy sea encroached from the southeast and reached the headwaters of the Tonto basin in central Arizona (Pl. 5b) where, at the present time, the Troy quartzite overlaps the rocks of the pre-Cambrian Apache group and abuts against the Mazatzal quartzite. In southeastern Arizona the Troy quartzite likewise rests either on the Apache group or on the post-Apache effusives. The upper part of the Troy, often referred to as the Bolsa quartzite, overlaps the Troy quartzite proper on the pre-Algonkian platform composed of schist and granite, thus outlining the greatest extent of the Troy basin.

The northern boundaries of all consecutively younger Cambrian formations are regressive to the southeast, marking a gradual retreat of the Cambrian sea which was accompanied by the successive introduction of younger faunas. The late Middle Cambrian shaly Santa Catalina formation has not been observed north of the confluence of the Gila and San Pedro rivers (Pl. 5c), and the somewhat younger Cochise formation with limestone beds which contain the *Olenoides pugio* [1] and *Eldoradia* faunas is even more restricted to the southeast and is not known far north of the Santa Catalina and Little Dragoon Mountains. Since these formations conformably overlie the Troy quartzite, the age of the latter is placed somewhere near *Elrathiella-Triplagnostus* time. (*Compare* Howell and Mason, 1938, p. 281.)

By the time of introduction of the *Tricrepicephalus texanus* and *Hesperaspis butleri* fauna, which characterizes the Upper Cambrian Abrigo formation, the sea had appreciably withdrawn farther southeastward, and the Abrigo formation is not found north of the Santa Catalina Mountains near Tucson (Pl. 5d). The strata with *Aphelaspis-Iddingsia*

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[1] *Neolenus intermedius pugio* Walcott (1910, p. 35, Pl. 6, fig 9). *See* Kobayashi (1935, p 154).

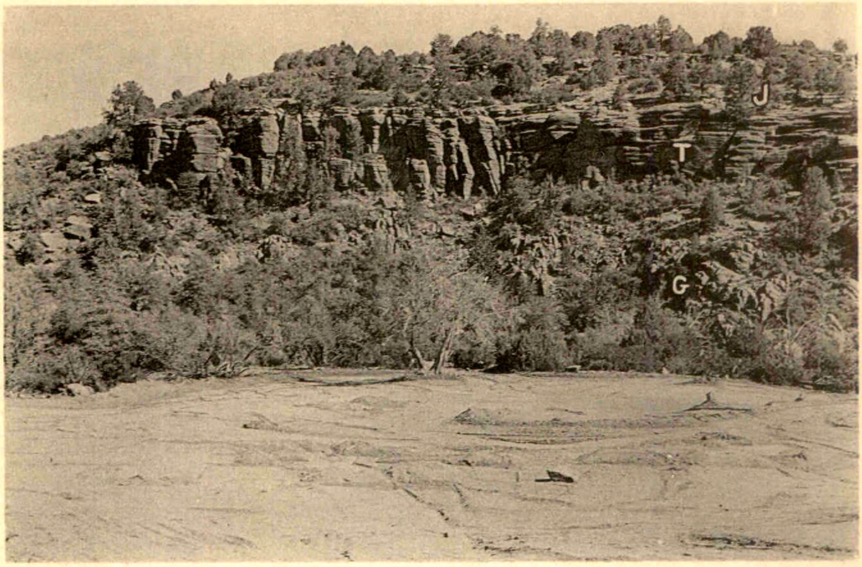

FIGURE 1. EAST VERDE CROSSING
Looking northwest. G, granite; T, Tapeats sandstone; J, Jerome formation with basal limestone resting directly on Cambrian.

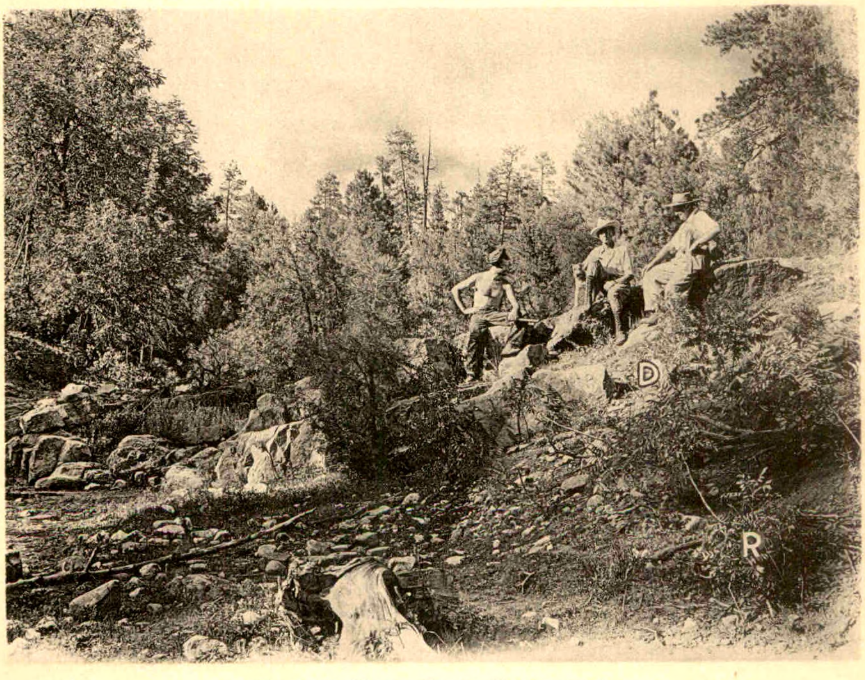

FIGURE 2. TONTO CREEK HEADWATERS
Looking north. D, base of Upper Devonian clastics with arthrodire fishes overlying R, Red Rock rhyolite.

PALEOZOIC STRATA ON PRE-CAMBRIAN ROCKS, CENTRAL ARIZONA

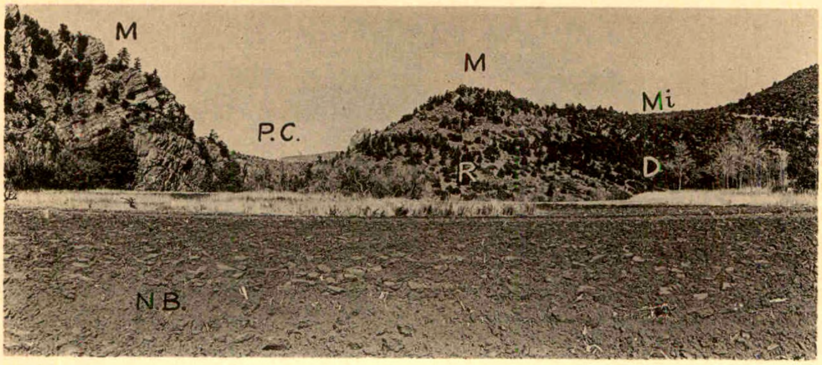

FIGURE 1. NATURAL BRIDGE
Looking north. Pine Creek, P.C., is deep in the gorge above and runs under the natural bridge, N.B., in foreground. R, Red Rock rhyolite; M, Mazatzal quartzite; D, basal Devonian sandstone; Mi, Mississippian Redwall limestone.

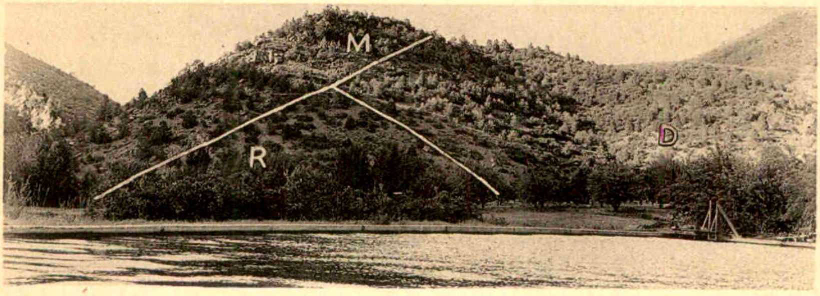

FIGURE 2. DETAIL OF FIGURE 1
East of Pine Creek. R, Red Rock rhyolite; M, Mazatzal quartzite; D, basal Devonian sandstone abutting against Algonkian rocks.

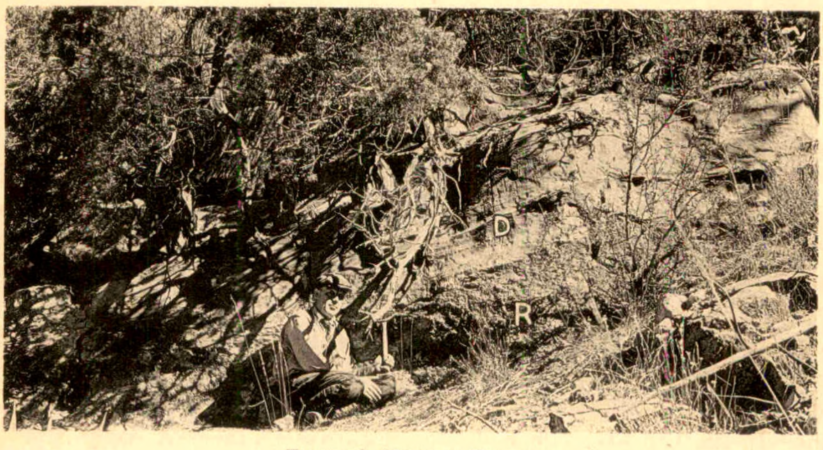

FIGURE 3. NATURAL BRIDGE
East side of Pine Creek gorge below settlement. D, basal Devonian sandstone on R, Red Rock rhyolite.

PALEOZOIC — PRE-CAMBRIAN CONTACTS AT NATURAL BRIDGE, CENTRAL ARIZONA

assemblages (Rincon limestone and Peppersauce Canyon sandstone) have approximately the same areal extent, whereas the beds with *Chariocephalus, Idahoia, Irvingella,* and *Maladia* (Copper Queen limestone) are restricted to the very southeastern corner of the State.

It is concluded from the foregoing discussion that in Cambrian time there was no communication between the northwestern and southeastern depositional areas of Arizona. The absence of the late Middle Cambrian and Upper Cambrian in northern Arizona, as seen from the sections in the Grand Canyon, rather supports this contention. Recently E. D. McKee located strata with a *Crepicephalus* fauna northwest of the Grand Canyon area in the Muddy Mountains, Nevada, and in the Virgin Mountains which are partly in northwestern Arizona (personal communication, September 12, 1941). The northwestern basin was in open communication with the Cordilleran geosyncline. On the other hand, the Cambrian southeastern basin was not directly connected with the Cordilleran sea across Arizona and southeastern California, but there is ample evidence that the Paleozoic seas of southeastern Arizona trough deepened southeastward into Mexico.

The presence of *Olenoides* and *Eldoradia* faunas in the late Middle Cambrian of southeastern Arizona needs an explanation since the nearest known locality with *Olenoides* is the House Range, Utah; the nearest localities with *Eldoradia* are the Eureka district, Nevada, and the Marble Mountains, California. Theoretically the intercommunication was possible either north of Arizona through the present Colorado Plateau via a Ouachita waterway east of Mazatzal Land or south of it through Sonora. Against the first possibility is the absence of Middle Cambrian strata in the Texan part of the Ouachita trough and the lack of evidence of the *Olenoides-Eldoradia* faunas in the Cambrian of New Mexico.

In his earlier interpretation the writer was inclined to regard the Middle Cambrian strata of the Marble Mountains near Cadiz, San Bernardino County, California (Hazzard and Crickmay, 1933, Map 1), as the southernmost termination of the Cordilleran sea and postulated a view (Stoyanow, 1936, p. 462) that Mazatzal Land might have extended from southwestern Arizona across southeastern California into the present area of Peninsular ranges. Important data on the stratigraphy of Sonora have been obtained recently by two able Mexican geologists, Isauro G. Gomez L. and L. Torres. The collected paleontological material has been brought to the University of Arizona for identification. Messrs. Gomez and Torres very kindly provided the writer with valuable field observations. In the Rio Magdalena Valley, northwestern Sonora, early Middle Cambrian strata with *Glossopleura, Alokistocare,* and *Anoria?* are present. This is the Arrojos formation, an equivalent of the Bright Angel shale of the Grand Canyon. The type locality, Arrojos Hills, south of Caborca, is 50 miles east of the Gulf of California and 400 miles southeast of Cadiz, California, where similar faunas are known (Darton, 1907, p. 470; Clark, 1921, p. 6; Resser, 1928, p. 10; Hazzard and Crickmay, 1933, p. 74; Mason, 1935, p. 97; Hazzard and Mason, 1936, p. 233). The entire Paleozoic trough of northwestern Sonora, as inferred from the distribution of studied outcrops, extends southeastward for an unknown distance and is separated from the trough of southeastern Arizona and northeastern Sonora by a wide belt of Archeozoic schist and granite which connects to the northwest with the similar rocks of southwestern Arizona. This is the Altar Headland of Mazatzal Land. The Arrojos formation consists of about 900 feet of alternating shale and limestone. It rests unconformably on the Jojoba formation which is made up of massive limestone beds with reefs of *Collenia* and is provisionally placed in the Algonkian. The Arrojos formation appears to be overlain by Mississippian limestone.

The significance of this paleontological find is that it establishes an early Middle Cambrian invasion in Sonora from the Cordilleran Sea or the Pacific Ocean as shown on an early map of Schuchert (1910, Pl. 52). The northwestern trend of this seaway was, however, to the west of that postulated by Schuchert if the localities with the *Glossopleura-Alokistocare* fauna in California, Nevada, and Arizona are considered. It is indicated as extending from Caborca, Sonora, through the Marble, Providence, and Nopah-Resting Springs Mountains in California, the Goodsprings quadrangle and the Muddy Mountains in Nevada, and northward along the Cordilleran geosyncline with an embayment or cul-de-sac into northern Arizona.

It is plausible to suppose that late Middle Cambrian waters entered southeastern Arizona through Sonora south of the Altar Headland. The *Glossopleura-Alokistocare* zone is about 200 feet above the base of the Arrojos formation, and younger faunas may be present within the latter. However, there is no evidence of the presence of the late Middle Cambrian *Olenoides* fauna like that of southeastern Arizona in any of the Cambrian localities of southeastern California.

Probably even in Late Cambrian time the southeastern Arizona basin did not belong entirely in the Ouachita trough (*compare* Schuchert and Dunbar, 1941, p. 126, Fig. 70) and a communication existed between the Cordilleran and Ouachita seaways through Sonora (*compare* Schuchert, 1923, p. 217, Fig. 5) with an arm into southeastern Arizona. The Upper Cambrian trilobite fauna of southeastern Arizona associated with and following *Tricrepicephalus texanus* has but a few common genera with that of Texas. *Chariocephalus, Hesperaspis,* and *Maladia* thus far are known only from Arizona, whereas *Burnetia, Conaspis, Pterocephalia,* and *Wilbernia* are found only in Texas; *Aphelaspis, Iddingsia, Irvingella, Tricrepicephalus,* and probably *Idahoia* (Bridge, 1937, p. 256) are common to the Upper Cambrian of both States.

DEVELOPMENT OF MAZATZAL LAND IN EARLY PALEOZOIC TIME AND LATE DEVONIAN TRANSGRESSION

The northwestern shore line of the southeastern Cambrian basin of Arizona lay progressively farther to the southeast as the sea retreated from Mazatzal Land. However, more of that land rose above the sea through the Ordovician, Silurian, and earlier Devonian. The only described Ordovician strata of Arizona, the Longfellow limestone of Beekmantown age (Lindgren, 1905, p. 65; Stoyanow, 1936, p. 483) in the Clifton-Morenci mining district and the unnamed limestone of the same age (Gilbert, 1875, p. 512; C. A. White, 1875, p. 77; Darton, 1925, p. 53; Stoyanow, 1936, p. 481) in the Dos Cabezas Mountains are both in southeastern Arizona near the New Mexico border (Pl. 5e). The exact age of the Coronado quartzite of the Clifton-Morenci area (Lindgren, 1905) is still unknown. It is reasonable to surmise, however, that the deposition of this quartzite did not correspond with the regressive phase of the Late Cambrian sea in southeastern Arizona. It probably marks the initial stage of the brief Early Ordovician transgression which eventually culminated in the formation of the fossiliferous Longfellow limestone.

West of this limited depositional area of Early Ordovician time there are neither Ordovician nor Silurian strata within Arizona, and none are definitely established in the Muddy Mountains, Nevada (Longwell, 1928, p. 22), in the Goodsprings quadrangle, Nevada (Hewett, 1931, p. 13), and south of Beaver County in Utah (B. S. Butler, 1920, p. 80, Pl. 4).

The general southward and eastward thinning of the Ordovician and Silurian formations in eastern Nevada and western Utah has been summarized in correlation tables prepared by Nolan (1935, p. 24). In an earlier paper Nolan (1928, p. 154) stressed the fact that the early Paleozoic sediments thicken westward toward the axis of the Cordilleran geosyncline from southwestern Utah and from Arizona—that is, from Mazatzal Land; on the attached map he marked the eastern limit of thick early Paleozoic sedimentation of the Cordilleran geosyncline and the western boundary of the pre-Carboniferous Paleozoic positive element, in this way outlining Mazatzal Land on the west.

In southeastern California Ordovician strata are present in the Nopah-Resting Springs Mountains (Hazzard, 1937, p. 322). The areas with Upper Ordovician limestone beds near La Casita in central Sonora (Dumble, 1900, p. 148; Schuchert, 1935, p. 135) are, according to R. E. King (1939, p. 1641, Pl. 1), between 28.30 N. Lat. and 29.00 N. Lat., east and west of 110.00 W. Long., about 150 miles south of the international border.

All available evidence indicates that the Ordovician and Silurian seas were in contact only with the extreme marginal regions of Mazatzal Land.

The Early and Middle Devonian waters bounded Mazatzal Land on the north and northwest considerably beyond the borders of Arizona, as is inferred from C. W. Merriam's (1940, p. 4) index map of better-known Devonian sections of western North America from the Inyo Mountains, California, northeasterly to Randolph, Utah. The presence of Middle Devonian fauna in the Goodsprings quadrangle, southeastern Nevada (Hewett, 1931, p. 15), and in the Nopah-Resting Mountains, southeastern California (Hazzard, 1937, p. 330), has not been definitely established. No Devonian strata have been reported from Sonora west of the Altar Headland either by previous or recent explorers.

The Late Devonian waters entered Arizona from the northwest and southeast, following in a general way the trend of Cambrian ingressions (Pl. 5f). Both in the extreme northwest and southeast of Arizona the basal Upper Devonian limestone rests directly on the underlying Cambrian strata with apparent conformity. This is not so in central Arizona where, especially in the Payson area between the headwaters of the East Verde River and the Tonto Creek, the Late Devonian waters, spreading from the northwest and the southeast, encountered a greatly eroded land made up of the pre-Cambrian schist, volcanics, Mazatzal quartzite, and granite. The granite, capped by the horizontal Cambrian Tapeats sandstone, occupied broad depressions, whereas the highlands were composed of the tilted Mazatzal quartzite and Red Rock rhyolite. Accordingly, the Devonian sea encroaching from the northwest deposited thin-bedded limestone directly and with apparent conformity on the Cambrian sandstone in the lowlands, but the elevated areas were vigorously attacked by the surf with the formation of basal conglomerates and coarse sandstones which contain remains of arthrodiran fishes. Later on, with further erosion and submergence, the clastic material was also carried into the depressions, and the basal limestone there was overlain by finer sandstone and shaly limestone, also with fish remains (Fig. 2). After this, with frequent but brief oscillations, the rest of the Jerome formation, culminating in coral reefs with brachiopod fauna, which extends from the East Verde headwaters northwestward beyond Jerome, was deposited (Stoyanow, 1936, p. 495).

Many incorrect interpretations have been made regarding the geology of this area, and deciphering of its paleogeography and stratigraphy is here briefly summarized.

Ransome (1916, p. 159) described a sequence of Paleozoic strata between Payson and Pine near Natural Bridge, Arizona, on Pine Creek, a confluent of the East Verde River. The nearly horizonal limestone beds, resting, as Ransome thought, on the Cambrian sandstone (Tapeats?), abut against the tilted Mazatzal quartzite underlain by a crystalline rock which Ransome took for granite but which Wilson later showed to be a volcanic rock, the Red Rock rhyolite. Ransome did not differentiate the limestone deposits which rest on the supposed Tapeats sandstone. He was impressed by the proximity of this series of nearly horizontal Paleozoic strata to the upturned pre-Cambrian Mazatzal quartzite and also by the fact that certain Paleozoic members contained angular fragments of this quartzite. He thus concluded that:

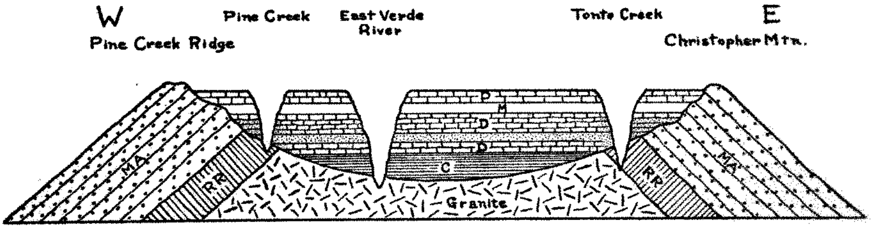

FIGURE 2.—*Sketch section across Payson Headland from Pine Creek Ridge to Christopher Mountain*

RR, Red Rock rhyolite; MA, Mazatzal quartzite; C, Cambrian; D, Devonian; M, Mississippian; P, Pennsylvanian. Faulting omitted. Distance 25 miles.

> "While it is difficult to understand the survival of a small sharp ridge as an isolated monadnock on a surface so nearly worn down to a plain as that over which the arenaceous deposits of the Cambrian sea were spread for thousands of square miles, it is still more difficult to account in any other way for this island-like mass of indurated tilted strata surrounded by dissimilar, nearly horizontal, and apparently younger beds."

Later E. D. Wilson (1922, p. 305) noticed that the basal Paleozoic sandstone in the Natural Bridge region differed lithologically from the Tapeats:

> "Due to the nature of the old uneven surface, the Tapeats sandstone is for the most part absent . . . however, there is usually between the pre-Cambrian rocks and the true 'Redwall' limestone a member which in different localities varies between limestone, sandstone, and arkose, that might sometimes be mistaken for the Tapeats formation. Whether this always represents merely the base of the 'Redwall' limestone, or sometimes the Tapeats sandstone as well, is difficult to say."

In the geological map of Arizona, published in 1923 by the Arizona Bureau of Mines in cooperation with the U. S. Geological Survey, the basal Paleozoic strata of Natural Bridge were mapped as Cambrian.

Darton (1925, p. 44, 249) followed Ransome in interpreting the same rock of Natural Bridge as the Tapeats sandstone.

In 1925 Schuchert and the writer found arthrodiran fishes between Payson and Pine, in a sandstone lithologically very similar to the Tapeats; they concluded that the entire sandstone might be Devonian (Stoyanow, 1926, p. 312).

Ten years later Hinds (1935, p. 28) referred to the basal conglomerate and sandstone of the Natural Bridge area as Tapeats and described some effects of erosion of the surface of Mazatzal Land that, in his opinion, had taken place in the pre-Tapeats time.

In 1936 the writer criticized Ransome's theory of monadnock in the upper part of the Pine Creek Canyon and explained the abutting of the horizontal Paleozoic limestones, which he had separated earlier (Stoyanow, 1926, p. 314) on faunal evidence into Devonian, Mississippian, and Pennsylvanian, by a fault (Stoyanow, 1936, p. 516-517). He discussed the fish-bearing deposits of the Payson-Pine area and their lithological relation to the Tapeats sandstone but placed the basal Paleozoic beds of Natural Bridge in the Cambrian in accordance with the views of Ransome, Darton, and Hinds.

Only in recent years has the interrelation of the strata in the headwaters of the East Verde River and the Tonto Creek come to light. The basal platform of the Payson area, referred to in this paper as the Payson Headland, is made up of schist and granite. On the north, the granite of this headland is overlain by almost horizontal Tapeats sandstone on which conformably rest the thin-bedded limestones, the basal members of the Upper Devonian Jerome formation (Stoyanow, 1936, p. 495-500) (Pl. 1, fig. 1). The outcrops of the Mazatzal quartzite at Natural Bridge are at the western margin of the headland, while on its eastern margin, in the headwaters of the Tonto Creek, 25 miles away, are cliffs also made up of the greatly disturbed and tilted Mazatzal quartzite like those of Natural Bridge (Fig. 2).

In the Tonto Creek headwaters, the Mazatzal quartzite, Red Rock rhyolite, and in places also the granite are overlain by the same basal conglomerates and sandstones, composed partly of chunks and boulders of the underlying rocks, as are present at the base of the Paleozoic sequence at Natural Bridge. In these basal beds the writer found, in the summer of 1939, abundant remains of arthrodiran fishes (Pl. 1, fig. 2).

Another reconnaissance in the Natural Bridge area revealed that independently of whether the contact between the Mazatzal quartzite and the abutting Devonian limestone on the east side of Pine Creek—the original location of Ransome's studies discussed above—might have been caused by local faulting (Pl. 2, figs. 1, 2), on the west side of the creek the basal conglomerates and sandstones, conformably underlying the younger Devonian strata, unmistakably rest upon the tilted Mazatzal quartzite (Fig. 3; Pl. 3, figs. 1, 2). On the east side of the creek, but

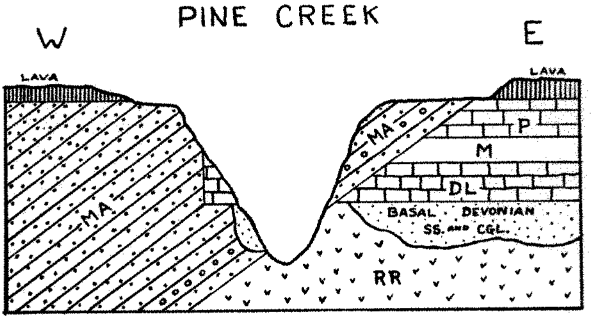

FIGURE 3.—*Sketch section across Pine Creek above the Natural Bridge settlement*

RR, Red Rock rhyolite; MA, Mazatzal quartzite; DL, Devonian limestone; M, Mississippian; P, Pennsylvanian.

below the Natural Bridge settlement, the same basal conglomerates rest on the rhyolite (Pl. 2, fig. 3).

These studies prove that the coarse basal conglomerates with angular fragments of the Mazatzal quartzite, including the large boulders of talus and fan debris described by Hinds (1935, p. 28) and others on Pine Creek (Pl. 3, fig. 3), are Upper Devonian and not Cambrian, and the accompanying erosion on Mazatzal Land, therefore, was coincident with the Late Devonian and not with the Cambrian transgression.

The basal Devonian clastics, where they rest on the Mazatzal quartzite, differ markedly from the Tapeats sandstone which is slabby whether it overlies the granite, as in the Payson Headland, or rests on the greenstone and schist, as it does farther northwest in the Jerome-Del Rio area.

As stated above, the basal complex of the central part of the Payson Headland is schist and granite. On the west, southwest, and east are isolated masses of the Mazatzal quartzite. The eastern margin of the headland most effectively separates the rocks of the southeastern area of deposition from those of the northwestern basin. This is the location of Ransome's barrier or "submerged ridge" which prevented the northwestward advance of the Apache and Troy waters. On the east, east of the marginal Christopher Mountain massive made up of the Mazatzal quartzite, in the deep canyons cut by Gordon and Haigler creeks, tributaries of Tonto Creek, the unconformable contact between the Mazatzal quartzite and the beds of the Apache group and the Troy quartzite is seen. These contacts were originally described by Ransome (1915, p. 116; 1916, p. 157) and later by E. D. Wilson (1922, p. 307; 1939, p. 1148) and by Darton (1925, p. 234).

The pre-Mazatzal and Mazatzal rocks of central Arizona, which in late Algonkian time stopped the Apache waters on the northwest, limited the Tapeats sea on the southeast and the Troy basin on the northwest in Cambrian time, formed, together with the overlying Cambrian strata, a headland which was attacked and encroached upon by the seas both from the northwest and from the southeast in Devonian time.

The exact nature of this Devonian headland in the northeastern direction, beneath the Colorado Plateau, is, of course, unknown, but, since the lithology, stratigraphic sequence, and marine faunas of the Jerome formation northwest of it are essentially different from those of the Martin formation to the southeast, communication between the two depositional areas was probably considerably impeded. Along the marginal, predominantly arenaceous, facies, however, the arthrodiran fishes existed from Superior to Jerome, and from there, in a more calcareous facies, to Flagstaff (Pl. 5f).

The Devonian on Tonto Creek is almost entirely restricted to the fish-bearing clastics which rest on the Mazatzal quartzite and are overlain directly by the Mississippian Redwall limestone. The *Cladopora* reefs with brachiopod fauna of the Jerome formation, which extend from Payson to Jerome and some distance north of it, are not developed on Tonto Creek.

In concluding this brief outline of the shaping of the Payson Headland and the relative role played in it by the basal Cambrian and Devonian strata, the writer wishes to comment on a tendency in recent literature to discredit in their original meaning the terms "Tonto group" and "Tapeats sandstone," which are extensively used in the Cambrian stratigraphy of northern Arizona. In an article entitled *Tonto group of Arizona is what? and where?* C. R. Keyes (1940, p. 237) placed the Tonto Creek near Fort Apache, in Navajo County, over 75 miles away from Gilbert's (1875, p. 60) type locality in the true Tonto basin, which is in the western part of Gila County, and claimed that the Apache group of Ransome is the original Tonto group. Keyes' assumption is based on Marvine's erroneous reference to the quartzites of the Apache group and the Troy quartzite in the upper basin of the Salt River as the "Tonto sandstone" (Marvine, 1875, p. 219). It is true, although not mentioned by Keyes, that the Tapeats sandstone is not present anywhere in the Tonto Creek canyon proper, and if Gilbert or Marvine had in view the sandstone exposed on that creek it was the Devonian basal sandstone with arthrodiran fishes discussed above. However, since Gilbert origi-

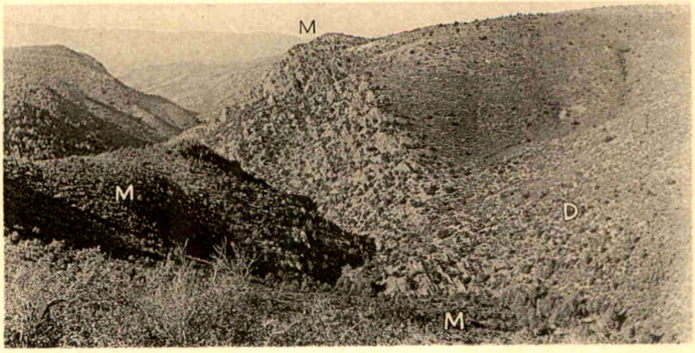

FIGURE 1. PINE CREEK GORGE ABOVE NATURAL BRIDGE
Looking south. D, Devonian sandstone resting unconformably upon M, Mazatzal quartzite, on the west side of Pine Creek.

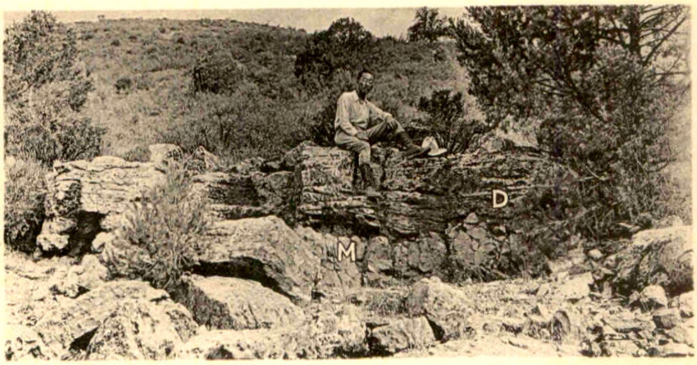

FIGURE 2. WEST SIDE OF PINE CREEK GORGE ABOVE NATURAL BRIDGE
Looking south. D, basal Devonian sandstone unconformably overlying M, Mazatzal quartzite.

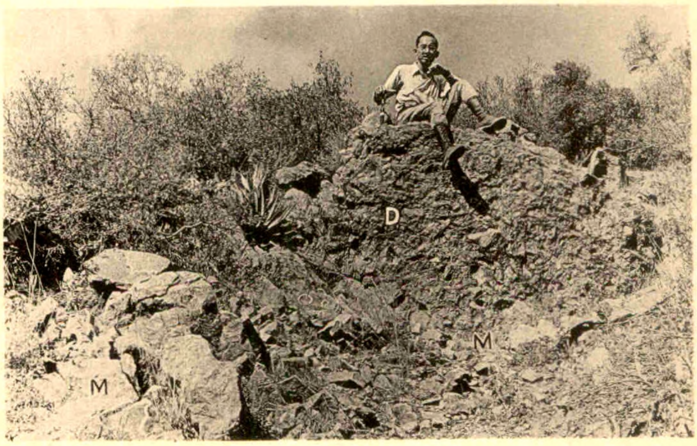

FIGURE 3. WEST SIDE OF PINE CREEK GORGE ABOVE NATURAL BRIDGE
Looking south. D, conglomeratic accumulations of basal Devonian material in pockets of M, Mazatzal quartzite.

CONTACTS BETWEEN DEVONIAN CLASTICS AND MAZATZAL QUARTZITE

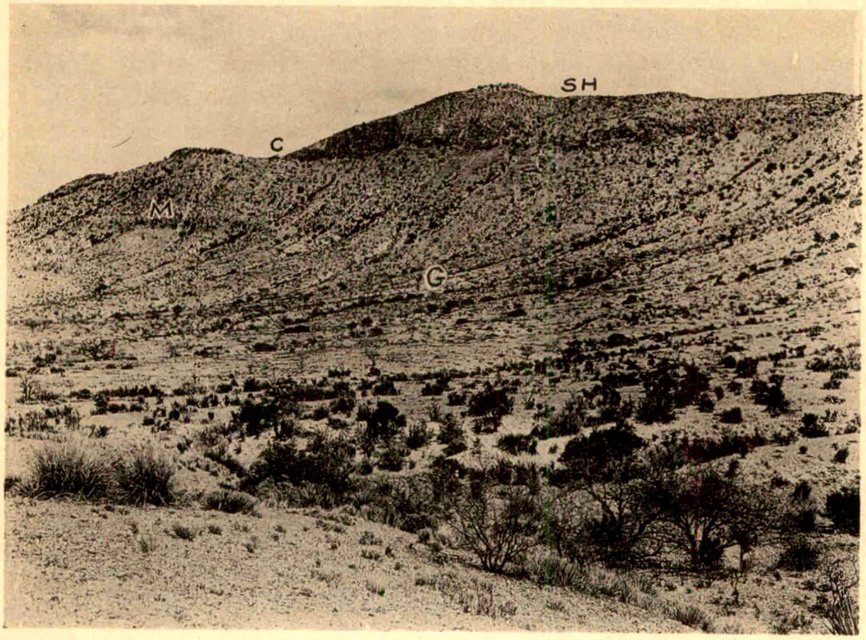

FIGURE 1. EAGLE BLUFF IN THE EMPIRE MOUNTAINS SOUTHEAST OF TUCSON
Looking northeast. G, gypsiferous series; M, basal limestone with Manzano fauna; C, contact between the Manzano beds and the Snyder Hill formation; SH, Snyder Hill formation.

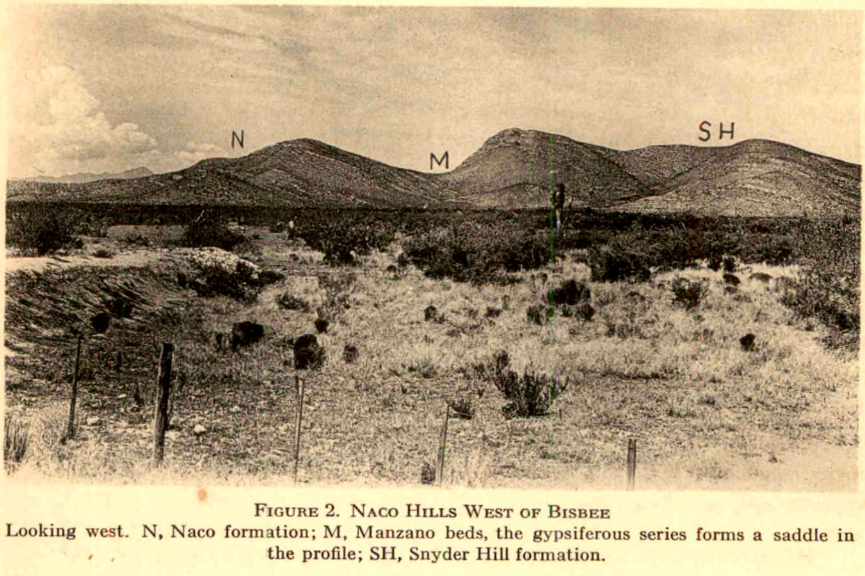

FIGURE 2. NACO HILLS WEST OF BISBEE
Looking west. N, Naco formation; M, Manzano beds, the gypsiferous series forms a saddle in the profile; SH, Snyder Hill formation.

PENNSYLVANIAN — PERMIAN SEQUENCE, SOUTHEASTERN ARIZONA nally referred to his "Tonto group" as located in the headwaters of the Verde and Tonto creeks, and because the western tributaries in the headwaters of Tonto Creek cut through Diamond Mesa and other mesas in which the Tapeats sandstone is well exposed, the presence of this sandstone, the basal member of the Tonto group, in the basin is beyond question. The fact that Marvine later confused the Apache beds with the Tonto group, or placed these beds within it, cannot invalidate the correctness of Gilbert's earlier definition.

In the northwestern basin the Devonian strata thin rather gradually in the northwestern direction from the Payson Headland as seen from the measurements at Jerome, Yavapai County (500 feet), and at Peach Springs, Mohave County (370 feet), but northeast of the axis of the basin these strata rapidly thin to 150 feet at Flagstaff, are not over 100 feet thick in the Grand Canyon area, and probably disappear northeast of the Grand Canyon.

The latest Upper Devonian strata, although different lithologically and faunally in the two depositional areas, indicate shallower waters. In the southeastern basin they are represented by the Lower Ouray formation, with *Camarotoechia endlichi,* which contains shale and sandstone in addition to limestone. This formation is known from the Little Dragoon Mountains near Benson to the upper basin of the Salt River, but it is absent in the Bisbee area in southeastern Arizona, where throughout the Paleozoic, but especially during the Late Devonian and Early Mississippian, the southeastern trough was deepest, and where the Devonian-Mississippian sequence consists, with slight local variations, only of limestone. In the northwestern basin, the Island Mesa beds likewise contain more shaly and arenaceous material than the underlying Jerome formation. These beds, with still undescribed fauna of heavy-shelled lamellibranchs and gastropods, are about 125 feet thick.

The small and isolated basin in the Clifton-Morenci area (Pl. 5f) is insufficiently known. Devonian age of its strata of argillaceous facies, the Morenci shale, was provisionally established by H. S. Williams on only one imperfectly preserved specimen of *Schizophoria* (Lindgren, 1905, p. 66-69). It is doubtful that this basin was directly connected with the southeastern trough.

The shallowing of the Devonian waterways toward the close of the period is clearly indicated all over Arizona except in the extreme northwestern and southeastern areas.

MISSISSIPPIAN SEAS AND PENNSYLVANIAN TRANSGRESSION

The Lower Mississippian strata of Arizona are well separated from the older rocks by a definite datum, the basal granular limestone which differs markedly from the underlying Cambrian and Devonian sedimentaries. The Escabrosa limestone of the southeastern basin and the Redwall limestone of the northwestern depositional area are uniformly granular, crystalline, and predominantly composed of crinoidal material (Ransome, 1904, p. 42; Noble, 1922, p. 54; Stoyanow, 1936, p. 505).

The development of the Lower Mississippian waterway in southeastern Arizona, which coincided with the introduction of new faunas, was the sequel to a rather rapid deepening of the former Late Devonian trough and not to a new encroachment of the sea upon Mazatzal Land (Pl. 5g).

In the Bisbee area, the Escabrosa limestone rests conformably on the unfossiliferous (?) upper limestone beds of the Martin formation and farther northwestward on the Lower Ouray formation. Nowhere is there any evidence of interruption in deposition, no "profound erosional unconformity at its base, one of the great peneplanal surfaces of the continent" as Keyes (1940, p. 237) interprets it. The influence of Mazatzal Land on the southeastern sea in Early Mississippian time is seen in the gradual thinning of the Escabrosa limestone from Bisbee toward the Payson Headland and in the appearance, in the approaches to the headland, of the red regolithic material from the pre-Mississippian rocks of Mazatzal Land in the limestone beds. The thickness of the Escabrosa limestone north of Douglas is 950 feet (Stoyanow, 1936, p. 511), 800 feet at Bisbee (Ransome, 1904, p. 43), 425 feet at Superior (Harshman, 1939), not much over 150 feet at Lake Roosevelt (Stoyanow, 1936, p. 508), and much less than that in the areas closer to the Payson Headland (Darton, 1925, p. 237).

The western limit of the southeastern basin is not clear. Gilluly (1937, p. 40) described a fanglomerate in the Ajo area in which among other components limestone material, carrying Late Paleozoic fossils known in southeastern Arizona, was observed. W. L. Thomas reports a large number of middle-sized limestone boulders containing crinoids scattered on the floor of Yuma Desert southeast of Yuma. The redeposited material of Ajo undoubtedly was derived from the sedimentaries of the southeastern basin. In northwestern Sonora, west of the Altar Headland, the Devonian is unknown, and the Mississippian strata, which rest directly on the Cambrian, contain beds with Lithostrotiontidae, not represented in the Mississippian faunas of Arizona but recorded in the Cordilleran seaway in California. To the east, the Modoc limestone of the Clifton-Morenci area (Lindgren, 1905, p. 69; Stoyanow, 1936, p. 511) was deposited in a small, isolated water body apparently not connected directly with the larger southeastern basin.

In the northwestern depositional area, the Early Mississippian waters spread over the Devonian strata in a northeastern direction as is seen from the overlap of the Redwall limestone on the Cambrian Muav limestone in the Grand Canyon where, according to McKee (1937, p. 341), the Redwall limestone thins eastward. The Redwall limestone in the Grand Canyon area is variously estimated as between 500 and 700 feet thick, depending on the interpretation of the age of the thinner limestone beds between the massive limestones of the Redwall and the base of the Supai formation. It is quite possible that, in the Grand Canyon or to the west of it, the Redwall limestone attains its maximum thickness within Arizona. Along the axis of the northwestern trough, the Redwall limestone thins to the southeast. In the Jerome area it is 300 feet thick, but it is only 80 feet thick farther southeast in the headwaters of the East Verde River (Stoyanow, 1936, p. 512, 513); in the upper basin of Tonto Creek the Redwall limestone has thinned to a few tens of feet. The earliest Lower Mississippian fauna of the southeastern depositional area is unknown in the Redwall limestone, from which it may be inferred that the Mississippian fauna, and probably the Mississippian waters, reached the northwestern basin somewhat later.

Toward the close of the Early Mississippian, the waters had completely withdrawn from the northwestern basin. In the southeastern basin, the withdrawal of the Mississippian sea was not so complete. In the southeasternmost part of Arizona the sea lingered through the Late Mississippian in a limited area in which the fossiliferous Paradise formation was laid down (Stoyanow, 1926; 1936; Hernon, 1935). This very limited area, in the Chiricahua Mountains north of Douglas, seems to have been the only part of Arizona under the sea in Late Mississippian time (Pl. 5h).

The marine cycle under discussion began with the Late Devonian transgression from the northwest and southeast toward central Arizona; the sea was deepest in Early Mississippian time; the recession of the Early Mississippian waters, which followed almost precisely in the same manner and direction as did the withdrawing Cambrian seas, left the entire State bare, with the exception of the lingering Late Mississippian water body in its southeastern corner.

During the following Pennsylvanian transgression, the sea, issuing again from the extreme southeast of Arizona, reached the Payson Headland and encroached upon it, but the Pennsylvanian sea did not spread far beyond the headland northwestward and did not occupy the northwestern area of deposition, the site of the previous seas in northwestern Arizona. Northwestern and northern Arizona was land through Pennsylvanian time with the possible exception of the extreme northwestern portion of the State in the neighborhood of the Grand Wash Cliffs and the Virgin River area (Longwell, 1921, p. 46; Reeside and Bassler, 1922, p. 66, 77; Longwell, 1928, p. 33; Longwell and Dunbar, 1936, p. 1203), since the evidence of Pennsylvanian faunas has been found in adjacent parts of Nevada and Utah. The Pennsylvanian water body of that region, however, belonged to a sea, the greater part of which was beyond the limits of Arizona and which at no time was in direct communication with the southeastern-central Pennsylvanian basin discussed in this paper (Pl. 5i).

In the extreme southeastern part of Arizona, the basal Pennsylvanian beds, like the basal beds of the Upper Devonian and the Lower Mississippian, are made up of limestone. They rest conformably on the Upper Mississippian Paradise formation in the Chiricahua Mountains, and with apparent conformity on the Lower Mississippian Escabrosa limestone at Bisbee. To the northwest, the first appearance of clastic rocks—intraformational conglomerates—within the basal limestone beds of the Naco, the Pennsylvanian formation of southeastern Arizona, has been recorded in the Empire Mountains about 45 miles northwest of Bisbee and 28 miles southeast of Tucson. At Superior, farther northwest, a well-defined conglomerate separates the Lower Mississippian strata from the basal Pennsylvanian limestones (Harshman, 1939). In the Payson area of central Arizona, the encroachment of the Pennsylvanian sea on Mazatzal Land is in full evidence. Basal and intraformational conglomerates contain boulders of the Mazatzal quartzite, thus indicating the progress of the invasion, and the distribution of Pennsylvanian strata shows that the sea had overcome the eastern barrier on the Payson Headland which was insurmountable for the Apache, Cambrian, Devonian, and Mississippian waters. The Pennsylvanian sea had spread over the Payson Headland to be checked only by the second line of the more resistant pre-Cambrian rocks at its western margin. The lithology of the Pennsylvanian strata on the headland is strikingly different from that in the areas southeast of it. Clastic rocks predominate. The thin, slabby, fossiliferous limestone beds in the lower part of the Pennsylvanian sequence are underlain and overlain by maroon shale, and in its upper part are brown sandstones and grayish thin-bedded and laminated shales of considerable thickness, as seen from the outcrops in the Tonto Creek headwaters. Such rocks are unknown in the Naco and Galiuro, the Pennsylvanian of southeastern and central Arizona.

LAND AND SEA IN PERMIAN TIME

The writer has shown that the southeastern basin was deepest in its extreme southeastern end through each of the outlined Paleozoic invasions upon Mazatzal Land. Within southeastern Arizona, every transgression started from the southeast, and during the withdrawals the waters either stayed there longer or lingered until the time for a new northwestward advance arrived. This was the area of the greatest accumulation of Paleozoic limestone deposits in the State.

The northwestern trough was permanently destroyed with the close of the Early Mississippian, and no broad marine bodies occupied northern Arizona through the Pennsylvanian and the greater part of Permian time until the development of the Kaibab facies there at the termination of the Permian submergence in Arizona.

In southeastern Arizona, however, deeper-sea deposition went on from the Pennsylvanian into the Permian with only brief interruptions.

In pre-Kaibab time, and especially during the deposition of the earlier Supai, the broad, shallow brackish and fresh-water bodies of northern Arizona were controlled by the southeastern Permian sea, but the communication between the southeastern and northern regions was not through central Arizona. From time to time the marine waters of the southeastern basin reached, by brief incursions through eastern Arizona, east of Mazatzal Land, the shallow depressions where the deposition of the Supai red beds was going on (Pl. 5j).

The writer (1936, p. 521, 530) discussed the present status of the Upper Carboniferous stratigraphy in southeastern Arizona. When establishing the Naco limestone as a stratigraphic unit, Ransome (1904, p. 54) regarded it as prevalently Pennsylvanian. Girty (Ransome, 1904) tentatively separated the Naco into a lower part with well-known Pennsylvanian fauna and an upper part with a different assemblage of fossils predominantly composed of gastropods. It was pointed out by the writer that: (1) Within the total thickness of 3000 feet of Ransome's Naco, as seen in the Whetstone Mountains between Bisbee and Tucson, the Manzano fauna, as previously described by Girty (1909) from New Mexico and indicated for the upper part of the Naco by Darton (1926, p. 821), is present; (2) the gastropod fauna mentioned by Girty from Ransome's collections belongs in the Snyder Hill, a younger formation than the Manzano; and (3) the beds of Manzano age extend from southeastern into east-central Arizona, and limestone layers with Manzano fauna, the Fort Apache limestone, are found within the Supai formation near Fort Apache in the upper basin of the Salt River (Darton, 1925, p. 85; Stoyanow, 1936, p. 533). Consequently, the term "Naco" was restricted to the Pennsylvanian part of the original Naco limestone, and the presence of marine Permian strata in the southeastern basin was recognized.

In 1937 B. S. Butler and G. P. Sopp (1940) discovered a gypsiferous series in the Empire Mountains east of Tucson above the Pennsylvanian Naco limestone and below the Snyder Hill formation (Pl. 4, fig. 1). The presence of gypsum below the Snyder Hill formation south of Tucson was mentioned before by C. F. Park (1929) and E. B. Eckel (1930). In the summer of 1940, the writer located the gypsiferous series and limestones with the Manzano fauna between the Naco and the Snyder Hill in the Naco Hills near Bisbee, the type locality of Ransome's Naco limestone. The Naco formation, rich in typical Pennsylvanian brachiopod and coral faunas, terminates in red shales and *Fusulina* limestones. The fusulinid specimens were submitted for examination to M L. Thompson who kindly informed the writer (personal communication, March 6, 1941) that they are all referable to highly developed representatives of *Triticites* of upper Pennsylvanian rather than Permian aspect They are comparable to the forms found in a zone at the top of the Magdalena in New Mexico. Above the Naco a sequence of shales, gypsum beds, and thin-bedded limestones aggregate 200 feet. These much softer beds form a saddle in the profile of the Naco Hills (Pl. 4, fig. 2). Within the upper part of these strata are limestone beds with the Manzano fauna. The term "Manzano" is applied here to the two discussed units in a broad sense, implying strata with the Permian Manzano fauna as described by Girty rather than any accepted or suggested divisions of the Manzano group in New Mexico. The Snyder Hill formation, one of the most resistant of the Arizona Paleozoic rocks and often the cap rock of mesas and buttes in southeastern Arizona, is at the top of Ransome's original "Naco limestone"

In a northwesterly direction, toward Tucson, the Permian basin was somewhat shallower in Manzano time; there are clastic rocks at the base of the gypsiferous series; the gypsum layers are more abundant and thicker; and the fossiliferous limestones in the upper part of the sections alternate with quartzites. These strata, between the Naco and the Snyder Hill, are now known in the Whetstone Mountains, Cochise County, and in the Empire Mountains, Pima County, southeast of Tucson; in the Santa Rita Mountains and in isolated hills between Tucson and Nogales, south of Tucson. In the Empire Mountains the gypsiferous series is 1460 feet thick, and the overlying fossiliferous limestones and quartzites are about 500 feet thick (measured by F. W. Galbraith). The gypsiferous beds in the Santa Rita Mountains are 1300 feet thick; the limestones with the Manzano fauna aggregate 900 feet (measured by W R. Jones). The Manzano fauna is found invariably in the basal beds of the limestone series. It differs markedly from the faunal assemblages of the underlying Naco formation. The well-preserved silicified fossils of numerous, predominantly small shells of gastropods and lamellibranchs are easily leached out of limestone. Productids are exceedingly scarce in this facies The most abundant diagnostic species are: *Orthonema* *socorroense* Girty, *Phanerotrema manzanicum* Girty, *Nucula levatiformis* Walcott, *Astartella subquadrata* Girty, and *Schizodus wheeleri* Swallow? Girty, found with a number of closely allied forms. Farther north, in White River Indian Reservation, the Fort Apache limestone, with a similar Manzano fauna, rests on the Kinishba beds with a Permian flora, corroborating the Permian age of this molluscan fauna.

The Snyder Hill formation, with a novel gastropod fauna, apparently has a much more limited northward extension (Pl. 5k). With the gastropods occur dictyoclosti of the *ivesi-bassi* group, although not identical with the species from the Kaibab limestone, and also the characteristic forms of the *occidentalis* and *indicus* groups. The writer (1936, p. 532) separated under the name Chiricahua limestone higher beds in the extreme southeastern part of Arizona in which the Kaibab species, like *Aviculopecten coloradoensis* (Newberry) and *Dictyoclostus bassi* (McKee), are found. These strata, in which the presence of *Waagenoconcha montpelierensis* (Girty) has recently been established, are now known to extend west of Tucson. This unit is comparable to the Kaibab limestone, *sensu lato,* inclusive to the *Waagenoconcha montpelierensis* zone. Unlike the Kaibab limestone, however, the Snyder Hill formation and the Chiricahua limestone contain exotic species so far unknown in the Permian of the Southwest. Such, for instance, are a species of *Trachydomia* which is close to, though not identical with, *T. magna* Hayasaka (1938) from the Permian of Japan, and *Meekella* sp. of the group of *Meekella kueichowensis* Huang (1933, p. 31, Pl. 4, figs. 8-9) from the Permian of southwestern China.

The "Kaibab limestone" is heterogeneous and was deposited under oscillatory conditions which prevailed in northern Arizona during Permian time, as may be inferred from the rapid lateral and vertical variations of its members. Recently it has been divided by McKee (1938) into a lower unit, the Toroweap formation, and an upper unit, the Kaibab formation proper. This separation was established on the basis of an unconformity between the two formations and also on the vertical distribution of two species of dictyoclosti of the *Dictyoclostus ivesi-bassi* group. According to McKee's earlier interpretation (Stoyanow, 1936, p. 528, 529) *Dictyoclostus bassi* characterized the lower division, whereas *D. ivesi* was characteristic of the strata above the unconformity. With the progress of his research McKee firmly established the fact that the stratigraphic range of these two species is exactly reversed, *D. ivesi* being restricted to the Toroweap, and *D. bassi* to the Kaibab formation proper.

As has been mentioned above, the Snyder Hill formation is sharply distinguished from the underlying Manzano by the sudden advance of dictyoclosti of the *D. ivesi* and *D. occidentalis* types. In the polyphyletic

Semireticulati of the Arizona Permian *D. ivesi* and *D. bassi* are closely allied, as was recognized by McKee. Although there is no appreciable segregation of smaller forms referable to *D. ivesi* in the lower division of the Snyder Hill, the larger forms identifiable with *D. bassi* predominate in the upper Permian sequence of southeastern Arizona.

As pointed out by McKee, the Manzano molluscan fauna in northern Arizona is pre-eminently developed very high in the marine facies of the Permian strata. Thus, for instance, *Nucula levatiformis* and *Schizodus wheeleri* occur in the two upper divisions of the Kaibab—that is, in and above the *Waagenoconcha montpelierensis* zone (McKee, 1938, p. 161, 162, 244). In southeastern Arizona there is no marked penetration of the Manzano fauna into the basal Snyder Hill, and if there is a recurrence of such a faunal assemblage anywhere within the Snyder Hill formation and the Chiricahua limestone it has not yet been discovered.

Both the limited areal extent of the Snyder Hill and the Chiricahua limestone and the composition of their faunas suggest no direct communication between the southeastern basin and northern Arizona east of Mazatzal Land at the close of the Permian in the State. McKee (1938, p. 173) thinks that, during the development of the open-sea facies of the Kaibab, a communication might have existed between the depositional areas of northern and southern Arizona through western Arizona. There is no definite corroboration of this postulation at present, but there are certain facts that deserve consideration. Darton (1925, p. 215-223) described outcrops of limestone, closely associated with schist and granite, in the Harquahala, Plomosa, and Buckskin Mountains in western Arizona. He found some Carboniferous fossils in the less altered limestone beds of the Harquahala Mountains. He also noticed Carboniferous fossils in the redeposited boulders south of those mountains. In 1939 the writer examined post-Permian strata in the Bill Baer Hills south of the Harquahala Mountains. The strata which contain the boulders are arkosic sandstones with some quartz and alternate with thin-bedded quartzites. At many horizons of the sandstone are numerous, large and small, very angular boulders of blue limestone. The size and angularity of the boulders argue against extensive transportation. The fossils are satisfactorily preserved, and the fauna consists of crinoids, sea urchins, bellerophontids, and productids. This fauna is undoubtedly Permian but does not clarify its relation either to the Snyder Hill or the Kaibab faunas. On Plate 5k the Harquahala area is outlined with question mark as a possible site of a Permian basin of unknown relation. It is interesting, however, that in California, 1 mile north of the International Border and about the same distance west of the Colorado River, well-preserved silicified specimens of *Dictyoclostus occidentalis* and other brachiopods of the Snyder Hill-Kaibab affinities are found in the lower beds of gravelly mesas along the All-American Canal survey line (R. Lenon, personal communication and specimens).

At present the stratigraphic position for an equivalent of the Coconino sandstone or of the Toroweap formation is not determinable within the continuous development of the open sea facies of southeastern Arizona represented by many hundreds of feet of an uninterrupted limestone sequence.

The relation of the Snyder Hill to the Kaibab formation also is not quite clear. McKee (1938, p. 35-54) has shown that each of the three members of the latter formation is composed of a number of localized facies. Facies 1 of the middle member rapidly wedges out south of Flagstaff. Facies 5 of the same member, with *Waagenoconcha montpelierensis*, is developed principally in Utah. In the upper member of the Kaibab the possible open-sea facies 4 is even more localized. For this reason no attempt is made here to date the Kaibab basin in relation to the seas of southeastern Arizona prior to the deposition of the Chiricahua limestone.

Plate 51 represents the rapidly decreasing Permian marine basins of Arizona toward the close of *W. montpelierensis* time. At that time the southeastern basin was in communication both with the Ouachita seaway and with the Pacific waters. I. G. Gomez and L. Torres (personal communication and specimens) found a Permian fauna with *W. montpelierensis* in Monos Hills (Monos formation), northwestern Sonora, about 50 miles east of the Gulf of California.

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SHIFTINGS OF THE MISSISSIPPI RIVER IN RELATION TO GLACIATION

BY FRANK LEVERETT

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ABSTRACT

Preglacial drainage of the present Mississippi watershed was in two distinct systems. A western system headed in southern Minnesota and flowed southeastward across northeastern Iowa to Muscatine, thence southward along the present course, except for a few miles opposite the lower rapids The eastern system from St. Paul, Minnesota, southward to the upper rapids, turned eastward at LeClaire, Iowa, and discharged through the lower part of the Illinois valley.

These drainage systems were displaced and shifted back and forth by ice sheets lying both east and west of the river. In the Nebraskan and Kansan stages disturbance by the Keewatin ice sheet initiated the cutting of a new valley from Muscatine eastward to the old eastern drainage. In the third or Illinoian stage the Labrador ice sheet covered the eastern system below Clinton, Iowa, and forced all of the drainage into a brief or temporary course through eastern Iowa outside the Illinoian ice border. When the Illinoian ice sheet melted drainage from eastern Iowa down the old valley south from Muscatine was reopened. The entire present Mississippi became established in early Wisconsin time, about 80,000 years ago. In late Wisconsin time it also carried the Glacial Lake Agassiz drainage.

INTRODUCTORY STATEMENT

In an early report, the writer (1899) discussed briefly the displacement of the Mississippi River between the upper and lower rapids by the Illinoian glaciation. The subject was little more than outlined, and did not fully interpret the length of time the river occupied the channel it opened outside the west border of the Illinoian drift; it merely stated that the river returned to the present course from Muscatine to the head of the lower rapids at Montrose when the ice sheet melted from the intervening district. That it returned rather promptly was suggested

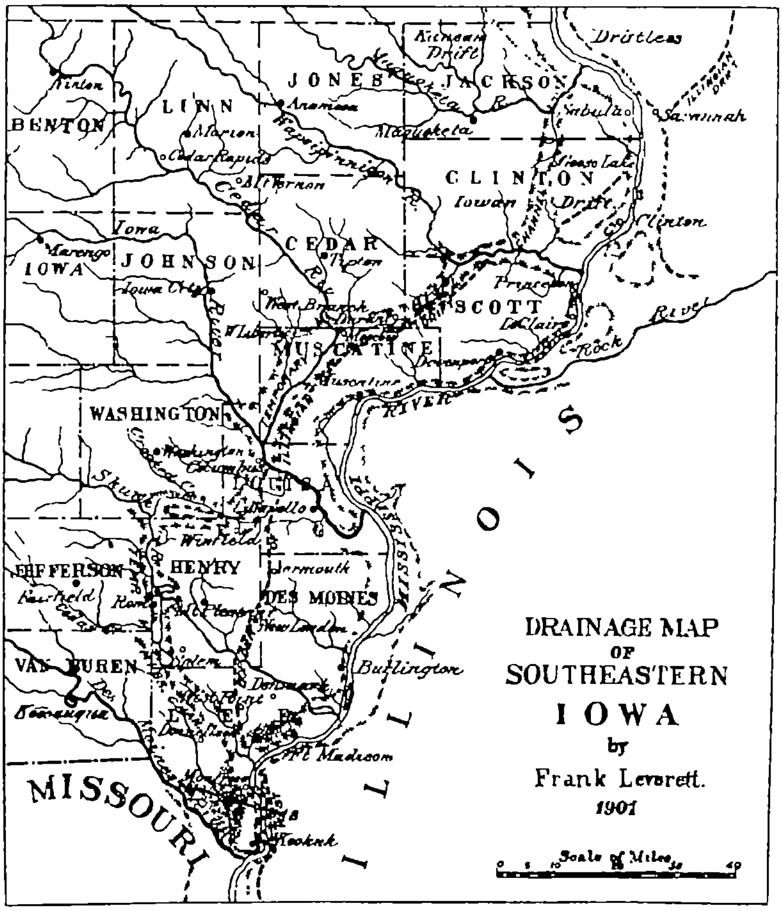

FIGURE 1—*Drainage map of southeastern Iowa*

Including the upper rapids at LeClaire, the lower rapids above Keokuk and features bordering the Temporary Mississippi channel in eastern Iowa From Leverett, 1901, p 40.

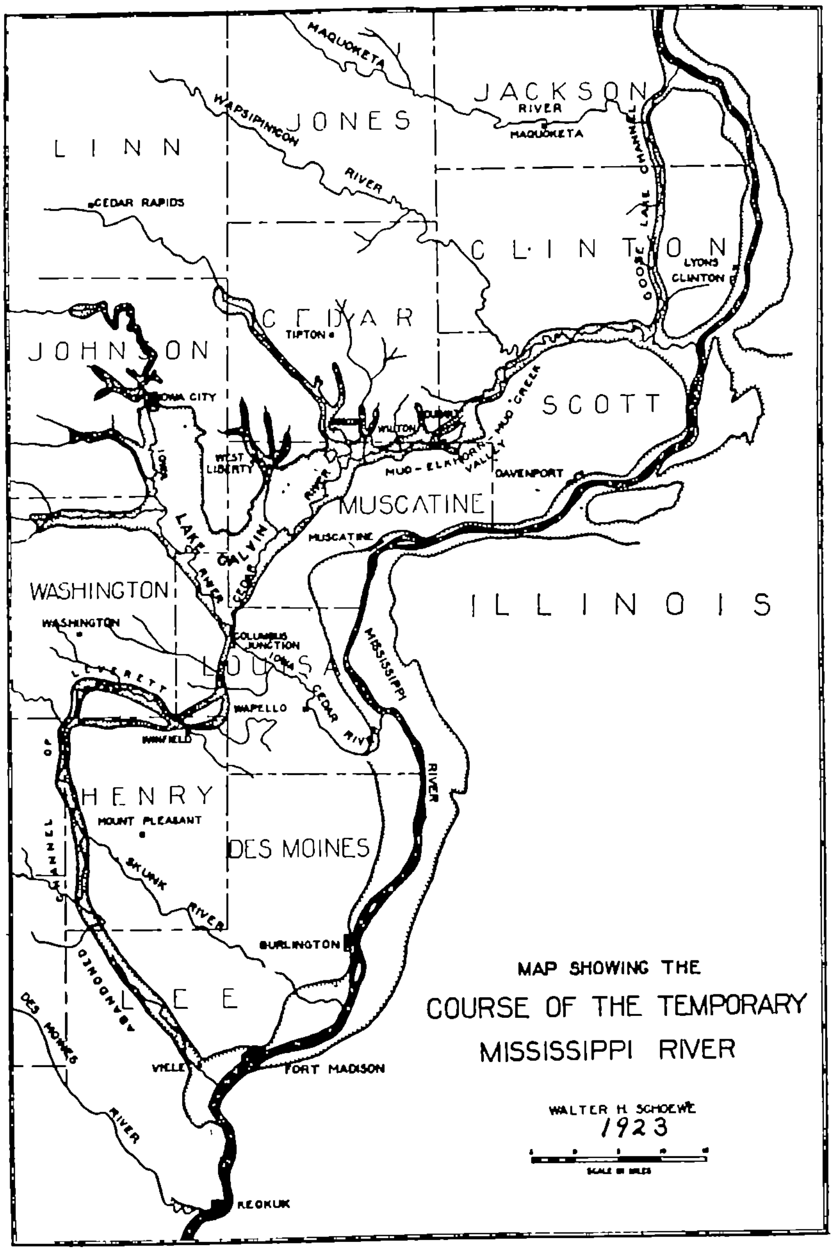

Figure 2—*The course of the Temporary Mississippi River*

From W H Schoewe, 1923, p 424

by the slight depth of the temporary channel, only 25-40 feet throughout much of its course. A long-continued stream of such volume should have done much more work, for it had a descent of more than a foot to a mile in the 75-80 miles of its course, about double the rate of the present river. It carried the drainage of the upper Mississippi as well as that from the Illinoian ice sheet. This temporary channel is shown in Figures 1 and 2.

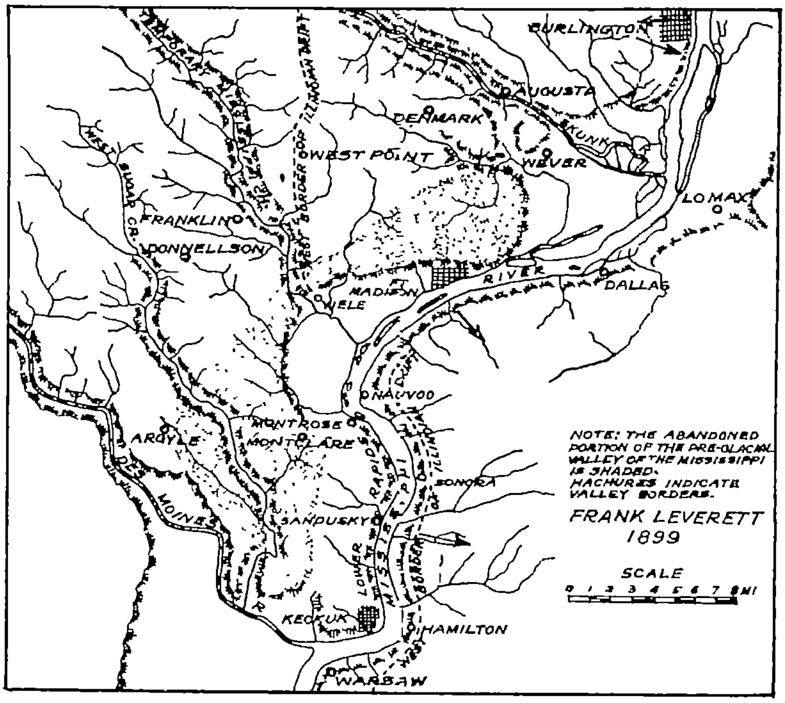

FIGURE 3.—*Sketch map of region around the lower rapids of Mississippi River*

Arrows indicate glacial striae. From Leverett, 1899, p. 468.

Supporting evidence of an early abandonment of this channel is found in the work done by such streams as Skunk River, which turned away from this channel to open a course across the Illinoian drift and join the re-established Mississippi River. Skunk River flowed for a considerable period in a shallow valley 20-30 feet deep, but nearly 2 miles wide, which is nearly double that of the Temporary Mississippi channel, although its volume was but a small fraction of that of the Mississippi River. The shallowness was an adjustment to the level at which the re-established Mississippi started to excavate a channel in the lower rapids, which was about 120 feet above the present level of the river at their head. Skunk River has since deepened its valley 100 feet or more in a narrow trench inside the limits of the broad, shallow valley, thus keeping pace with the lowering of the Mississippi channel across the rapids. It is of interest to note that this broad shallow valley turned south at Augusta, Iowa (Fig. 3), and headed toward the lower rapids, thus favoring the writer's interpretation of a prompt establishment of the Mississippi in the course through the rapids. Had these features been brought out more fully probably Schoewe (1924) would not have dissented from the writer's interpretation.

Figures 1 and 2, showing the course of the Mississippi in its relation to the Illinoian ice sheet, supplement each other. Figure 3 illustrates the features near the south end of the Temporary Mississippi channel, and at the lower rapids of the present river. The shallow valley of Skunk River is indicated by hachures. The filled preglacial Mississippi valley that passes around the rapids is also shown. Its size and the date of filling are discussed below.

A gumbotil of very moderate depth formed on the surface of the Illinoian drift in the early part of the somewhat lengthy period in which the river has been opening its channel through the lower rapids, a period estimated to cover about 160,000 years, and prior to the post-Sangamon loess deposition which preceded the culmination of the Labradorian-Wisconsin glaciation, estimated to have been some 80,000 years ago. The ordinary thickness of this gumbotil is 20 to 30 inches, or only one-third to one-half of the 5 feet reported by Schoewe. Very few exposures in this region have a thickness of 5 feet.

LAKE CALVIN

The name Lake Calvin was applied by Udden (1899, p. 350-356) to the lake that occupied a lowland in the lower course of Cedar River, 5 or 6 miles wide, which stood in the line of the Temporary Mississippi River. A silt deposit of this lake, exposed near Wilton, Iowa, contained the bones of a proboscidian which Calvin reported in 1874 to President Thatcher of the University of Iowa. He was unable to determine whether the bones were those of mastodon or elephant as the skull and teeth were missing. Calvin's early recognition of the silt as a lake deposit made the application of his name to the lake most fitting.

The level of Lake Calvin was determined by the altitude of the head of its spillway or outlet near Columbus Junction, which was put at 710 feet in the author's (1899) report on the Illinois Glacial Lobe.

In the 12 miles between there and Winfield it drops to 698 feet, corrected railway levels making it 5 feet less than stated in the former report. In the 25-30 miles from Winfield to Rome, on Skunk River, it drops to 670-675 feet. From there to the head of the lower rapids the rate is fully a foot to the mile. The channel is cut to a depth of only 25-40 feet below the bordering plain, and is less than 1½ miles wide. Its depth and breadth are so small that it has the appearance of a stream channel rather than a valley proper, the flood plain being the bordering upland.

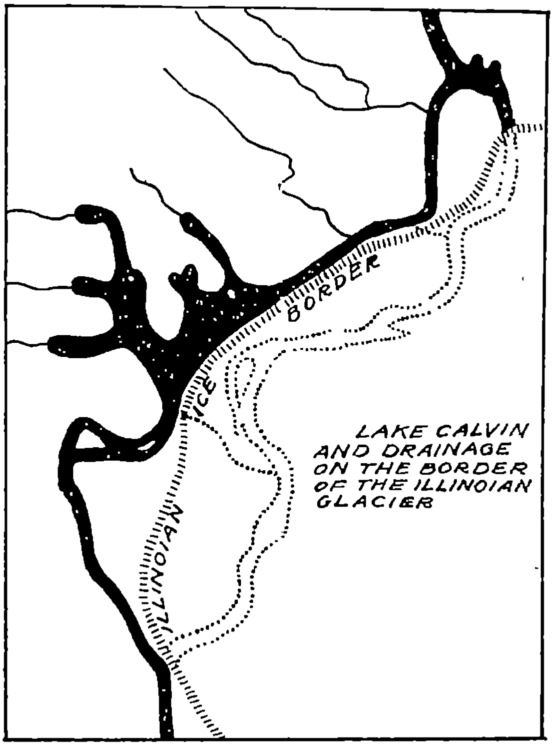

FIGURE 4—*Lake Calvin, and drainage on the border of the Illinoian ice sheet*

From Schoewe, 1924, Plate 17.

The level of Lake Calvin was a few feet higher than the bed in the head of the outlet, for its silt deposits cover ground 720 feet above sea level on the divide between the Cedar and Wapsipinnicon rivers near Durant, Iowa. As stated by Udden, any shore lines that the lake may have formed were too weak to be traceable after so long a lapse of time. With an altitude of 720 feet, the ponding must have extended up the Mississippi Valley far beyond the part that was covered by the Labradorian-Illinoian ice sheet, at least to the vicinity of Lake Pepin. It may there have received the drainage from the Patrician-Illinoian ice sheet. In this northern extension, as in Muscatine County the shore lines are probably obliterated. The depth of sediment on the bed of Lake Calvin appears also to have been very moderate, a fact that strongly favors the view that the lake was short-lived.

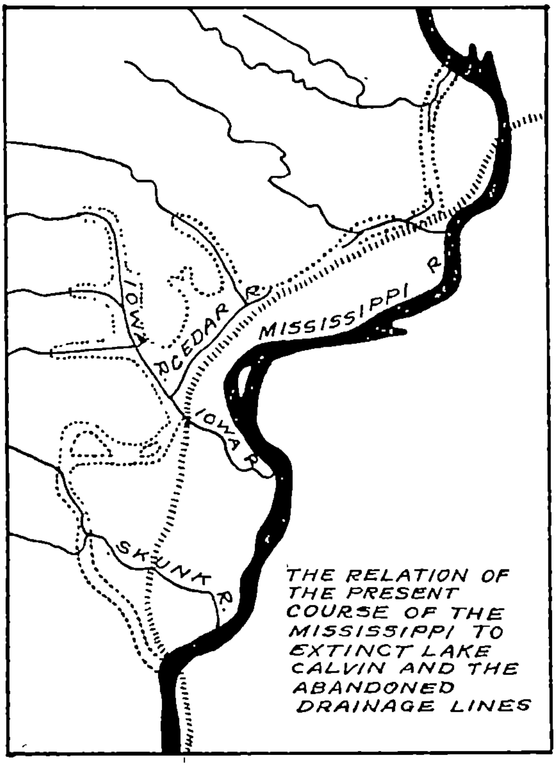
IOWA R.
CEDAR R.
MISSISSIPPI R.
IOWA R.
SKUNK R.
THE RELATION OF THE PRESENT COURSE OF THE MISSISSIPPI TO EXTINCT LAKE CALVIN AND THE ABANDONED DRAINAGE LINES

FIGURE 5—*Relation of the Mississippi River to Lake Calvin and abandoned drainage lines*
From Schoewe, 1924, Plate 18.

TIME NEEDED TO FORM THE TERMINAL MORAINE

The time required for the Illinoian ice sheet to build its terminal moraine may serve as a measure of the life of Lake Calvin, though the lake or ponded condition prevailed for a somewhat longer time—

partly before and partly after the building of the moraine, while the ice sheet advanced to and retreated from its terminal position. Where best developed the moraine is about 50 feet thick above the Kansan gumbotil and associated peat deposits remarkably preserved at its base. The building of such a moraine probably required centuries, and possibly a few thousand years. The advance to and retreat of the ice border across 20 miles intervening between the moraine and the present river seem likely to have taken a few more centuries. In the absence of more definite methods, such as it is hopd future students may work out in measuring rates of moraine building, we are only warranted in stating that at most the ice sheet occupied this district west of the Mississippi River only a very few thousand years.

WEST LIBERTY PLAIN AND OTHER FEATURES

The West Liberty Plain was described by Udden (1899). It occupies the bed of Lake Calvin in the Cedar River Lowland and at its northeast end, near Moscow, at 660 feet or about 60 feet lower than the neighboring Lake Calvin silts. He reports a downstream slope of about 3 feet per mile in Muscatine County to 615 feet at the line of Muscatine and Louisa counties.

Udden (1901) considered the terrace, 2 to 6 miles wide along the Iowa River in Louisa County, below the mouth of the Cedar River, a continuation of the West Liberty Plain. He does not clearly state the relation of this plain to Lake Calvin. However, the mention of the downstream slope of the plain and the inclusion of the Iowa River terrace in Louisa County in it, seem to indicate that he regarded it a fluvial feature, as to which the writer agrees.

Schoewe calls the West Liberty Plain of Udden the Intermediate Terrace since there is a higher terrace on Iowa River, at and below Iowa City, and also a lower one. He reports remnants of the higher terrace in the Cedar River drainage near West Liberty, Wilton, and along Cedar River above Moscow. It stands about 680 feet on Iowa River at Iowa City. It slopes about 16 inches per mile, as compared with 2 feet on the Intermediate Terrace, its altitude being 670 feet at Hills about 7 miles below Iowa City, and 660 feet 2 miles south of River Junction. It is entirely too low to find continuation over the spillway of the outlet of Lake Calvin south of Columbus Junction. The only line of continuation available appears to have been that which the Iowa River has taken. It seems to be the partially filled pre-Illinoian course of drainage. The Illinoian drift in the axis of that valley seems to have been little if any higher than 650 feet. A detailed study along the borders of the present Iowa River valley below the mouth of the Cedar River may reveal remnants of the high terrace, the broad terrace in the valley being, as suggested by Udden, a continuation of the West Liberty Plain (Schoewe's Intermediate Terrace).

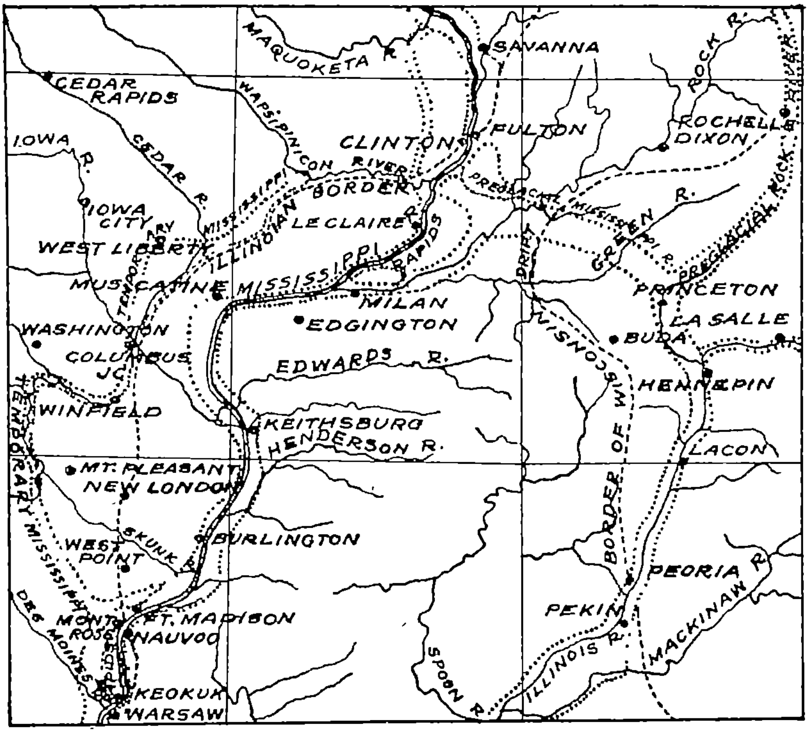

FIGURE 6.—*Map showing shiftings of Mississippi River*

Also the preglacial Rock River valley, the position of the Illinoian and Wisconsin ice borders, and the upper and lower rapids. From Leverett, 1921, p. 625

The Intermediate Terrace declines from about 650 feet at Iowa City to 615 feet at the Louisa County line, and to 606 feet at Fredonia, opposite the head of the Lake Calvin spillway (710 feet) near Columbus Junction, thus being fully 100 feet below that lake level. The High Terrace and the Intermediate Terrace are sandy, and not silty like the Lake Calvin deposits, except in a side valley near Wilton (Schoewe (1924), Figs. 25, 26, 27). In their make-up as well as in their downstream slope, they are fluvial rather than lacustrine. The Low Terrace, on the Iowa River below Iowa City, is of coarser material and is regarded by Schoewe as fluvial. The writer sees no reason for classing the others as lacustrine, or for continuing the life of Lake Calvin beyond the time its outlet through the Temporary Mississippi channel was in operation.

PRE-ILLINOIAN SHIFTINGS WITHIN THE MISSISSIPPI DRAINAGE BASIN

At the opening of the Glacial Epoch the upper Mississippi Valley above LeClaire, Iowa, and the valley of the present Mississippi below Muscatine, belonged to independent drainage systems. The upper Mississippi, and the Rock River drainage, flowed southeastward through the Green River Lowland to the Illinois at Hennepin, and thence down the Illinois Valley to its mouth (Fig 6). Another drainage system which appears to have headed in southern Minnesota flowed southeastward across northeastern Iowa to Muscatine. Its course and position have been only rudely determined from the well data of the region. The wells in its valley penetrate 200-300 feet of drift, while those outside the valley and main tributaries generally enter rock at less than 100 feet. More detailed collection and sifting of well records is needed to determine the course of the river and the limits of its drainage basin. A suitable name for it might then be given.

Below Muscatine, with the exception of a few miles near the lower rapids, this old valley is followed by the present Mississippi to the mouth of the Illinois River, where it connects with the eastern drainage from the upper Mississippi Valley. This combined stream, further augmented a few miles below by the Missouri River, utilized the large preglacial valley as far as Thebes, Illinois. It there partially abandoned the old valley and cut through a narrow peninsula to join the Ohio River. The valley is still utilized in flood stages below Thebes.

Warren (1878) reported the old or preglacial course around the lower rapids as indicated in Figure 3. Its valley is about 6 miles wide. Gordon (1890) reported a well record at Mont Clare, Iowa, which is within this abandoned part of the old valley. This shows its bed to be 126 feet lower than that of the present stream at the head of the rapids, or 374 feet above sea level. The filling of 305 feet is mainly Nebraskan drift. The overlying Kansan drift is only about 50-60 feet thick in that vicinity. The post-Kansan loess is but 6-12 feet thick. The Nebraskan drift is thus about 240 feet thick. The Paper Company well at Fort Madison, 11 miles up the valley, on a low terrace penetrated 145 feet of filling, mainly Nebraskan till, and entered rock at 379 feet above sea level. This is 123 feet below the low water level of the present stream at that point. The filling was sufficient at least by Kansan time, and very likely by Nebraskan time to permanently dislodge the stream from its preglacial course.

This seems a fitting time to give attention to the effect of the Kansan glaciation on drainage in part of the Mississippi Valley. It presents an interpretation held for some time but not hitherto published, and which is only vaguely expressed elsewhere in the literature.

The Kansan glaciation by the Keewatin ice sheet extended eastward several miles beyond the Mississippi River between Hannibal, Missouri, and Fort Madison, Iowa, and at least to the west bluff of the valley for 30 miles farther north, to Kingston, Iowa, where glacial striae trend southeastward. Striated ledges with this bearing are also found 10 miles east of the east bluff of the valley in the latitude of Quincy, Illinois. The position and bearing of the striated ledges near Hamilton and East Fort Madison, Illinois, and on the bluff 2 miles south of Burlington, Iowa, are indicated on Figure 3, as well as a later set of westward bearing striae at Burlington, referable to the Illinoian glaciation.

Evidently the displacement of drainage in this part of the Mississippi Valley was complete in Kansan time. Probably at this time some of the work of opening the valley eastward from Muscatine was enacted. Work probably started there in the preceding glacial stage, for the Nebraskan glaciation was similarly related to that part of the Mississippi Valley. However, that is an open question depending on whether the Nebraskan drift filling was high enough to prevent the Mississippi from returning to its pre-glacial course west of the lower rapids. In any case, the water from the melting Nebraskan ice sheet probably found eastward discharge there. How long the post Nebraskan drainage may have utilized this line is uncertain. As to the post-Kansan drainage through this line, very probably it persisted in the eastward course until the Illinoian ice invasion from the Labradorian Center; also the drainage of Iowa River may have utilized it. As to Skunk River, its present lower course is post-Illinoian. The pre-Illinoian course seems likely to have continued eastward from Brighton in southern Washington County, and it may have joined Iowa River. It can be affirmed with certainty that the Des Moines River basin drained southward in its present course after the Kansan glaciation.

The new section of the Mississippi Valley under consideration may have been much smaller than now when its eastward-flowing drainage gave place to the great stream now flowing westward. The size of the valley is well shown in the Edgington and Milan topographic maps of the U. S. Geological Survey. It averages about 1½ miles in width and has steep bluffs 150-200 feet high. Standing as it does within the district invaded by the Labradorian ice sheet in the Illinoian stage the valley was probably partially filled with Illinoian drift. On the bordering uplands this drift is only 20-30 feet deep. Unless the valley received a greater amount, the Mississippi River would have required only a relatively short time to remove it. The preglacial col or drainage divide in this valley was thought by Udden to have been near Fairport, Iowa; but the pointing of the present drainage on the borders of the channel seems to place it near Montpelier, 5 miles farther east. The upland there is fully as prominent as near Fairport.

CHANNELING AT THE UPPER RAPIDS

The author (Leverett, 1899) left open the question of the date of establishment of the Mississippi River in its course through the upper or LeClaire rapids. The work accomplished in eroding its trench is very slight compared with that involved in the trench at the lower rapids. The rock excavation is restricted to a short section, only 2 to 3 miles long, and nowhere more than 75 feet deep, the remainder being in loess and drift. However, the energies of the Mississippi River have been divided among several lines of discharge in that vicinity making it difficult to evaluate this part of its work. The broad Meredosia Slough, as may be seen by reference to the Cordova topographic map (Leverett, 1899, Pl. 18) is below 600 feet and is within the reach of present high stages of the Mississippi. At flood the stream joins Rock River and passes down its valley to the Mississippi, detouring the east side of the upland east of the upper rapids trench. The Cattail Slough makes a similar detour of the Mississippi flood waters a few miles farther north. It now stands above their reach because of accumulated peat and muck beds.

The Mississippi River seems to have divided its waters here for a considerable period in post-Illinoian time. Drainage may have been southeastward to the Illinois Valley through the Green River Basin (Fig. 6), about to the time of the culmination of the Wisconsin glaciation, when the strong moraines of early Wisconsin drift were formed. These have completely blocked the drainage from the upper Mississippi River to the Illinois Valley. It is estimated that this early Wisconsin culmination dates back about 80,000 years, or about half post-Illinoian time. The work at the upper rapids may thus be mainly within Wisconsin and post-Wisconsin time. Stream piracy by small tributaries of the Mississippi may have figured in opening this line of drainage, which has the advantage of being much shorter than the broad valleys that detour around the rapids.

This pre-Wisconsin drainage of the upper Mississippi to the Illinois Valley seems to require that the waters between the Illinoian and

Wisconsin glaciations that cut the lower rapids trench should involve mainly the district below Muscatine. The Cedar-Iowa drainage would then become the headwater line, with Skunk River the chief tributary, there being only small streams on the eastern or Illinois side. This interpretation may help to account for the slowness of the opening of the lower rapids trench.

VALLEY FILLING BELOW THE LOWER RAPIDS

The Mississippi Valley was filled at the lower end of the rapids during and after the Illinoian glaciation to about 90 feet above the present river, forming a terrace which at Warsaw, Illinois, stands a little above 560 feet. The Des Moines River built up its valley to a similar height at the "Yellow Banks" near its mouth. A section there, published by Gordon (1895, p. 243) is of especial interest in showing a succession of deposits, separated by two beds of black earth.

"YELLOW BANKS" SECTION

	Feet	Inches
(1) Clay, yellow, pebbleless (loess)	5	
(2) Silt, drab, pebbleless (loess)		15
(3) Earth, black, with a few small pebbles, apparently a flood-plain deposit	12	
(4) Clay, yellowish, local		6
(5) Sand, with a few small pebbles, with a layer of boulders at base	20–25	
(6) Earth, black, with yellow streaks, apparently an old flood-plain deposit	5–6	
(7) Gravel, with some sand, pebbles 2 inches or less in diameter	20	
(8) Clay, blue till, exposed to level of railway	15	
Total, about	80	

The upper bed of black earth, being at the base of the loess, seems to fall in Sangamon time and correlate with the Illinoian gumbotil. The thickness of 12 feet indicates a considerable time interval in harmony with the time involved in the weathering exhibited by the Illinoian gumbotil. The underlying pebbly sand, adjacent to that in the Warsaw terrace, connects with the filling from the excavation of the lower rapids, made at the mouth of the Des Moines The lower black earth and the gravel below it seem to be pre-Illinoian, and probably post-Kansan, while the blue till at the base is either Kansan or Nebraskan. The low altitude rather favors its reference to the Nebraskan glaciation, the top of the deposit being only about 500 feet above sea level. The surface of the Kansan drift in the upland back of the Yellow Banks, as shown by the Kahoka, Missouri-Iowa topographic map, is about 660 feet. Allowing 60 feet for the Kansan drift, the Nebraskan drift should reach about 600 feet, or 100 feet above its exposure at Yellow Banks.

At Keokuk, and at points above boulders are conspicuous 50 feet or more above the river. (*See* Leverett 1899, Pl. 7.) These probably are a concentrate from erosion of the Illinoian till where its edge was invaded in the development of the gorge. At Keokuk the boulders are under a sandy gravel the level of whose surface seems to correlate with the filling at Warsaw. Below Warsaw the altitude of the valley filling decreases much more rapidly than the fall of the present stream. Opposite Hannibal the filling at the mouth of Hadley Creek is only 35 feet above the river. The river falls only 25 feet in this interval while the terrace falls 90 feet.

PERMANENT ESTABLISHMENT OF THE MISSISSIPPI RIVER

The Mississippi River is in a new course above St. Paul. Winchell (1888), Grant (1890), and Sardeson (1908, 1916), measured post-Wisconsin time by recession of the waterfalls. The deeply-trenched valley at St. Paul is only a minor tributary of the preglacial upper Mississippi, the main line being farther west, beneath Lake Minnetonka. Its rock bed there is about 500 feet above sea level, or 180 feet below the level of the present river at St. Paul. Its course was southeastward through Dakota County and the St. Paul tributary entered it near Hastings. From Hastings, as far south as Clinton, Iowa, the river has a bedrock floor averaging fully 100 feet below the present stream. The width of the valley is nowhere less than 2 miles, and it averages about twice that width. As it is below the average width in the vicinity of Dubuque, Hershey (1897) suggested that the preglacial river headed only a short distance above that city. This view does not seem to find support in the slope of the rock bed, since no part of the floor slopes distinctly in the reverse direction as one approaches the head of the present stream. This continuous deep channel also seems to oppose a view held with some persistence by Trowbridge (1921) that this part of the Mississippi Valley is post-Nebraskan.

Just below St. Paul, at the Wisconsin drift border, the valley was remarkably aggraded with outwash gravel up to 940-960 feet—260-280 feet above the river. From there the surface of the outwash terrace descends fully 100 feet in the first 20 miles. It is only 790 feet above sea level at Red Wing and 125 feet above the river. At the mouth of Wisconsin River it is 75 feet above the Mississippi River, and 60 feet at Bagley, Wisconsin, 12 miles below. It then maintains a height of 50-60 feet above the river clear down to the head of the lower rapids.

The excavation of the lower rapids gorge may have been only down to this level when the Wisconsin filling reached there, thus helping to determine the height of the filling.

In the waning part of the Wisconsin stage the outlet of the great Glacial Lake Agassiz the "River Warren", carried a flood that swept away much of the Wisconsin filling of sandy gravel, and brought the valley about to its present condition. The flood widened the valley and cut back and greatly steepened the projecting headlands. The volume must have been far greater than in any of the modern high water stages. The lake lasted but a few thousand years and ended soon after its outlet was diverted to Hudson Bay. The entire life of the lake probably was not more than 10,000 years, its final 1000 years being tributary to Hudson Bay.

SUMMARY

Summarizing the succession of drainage shiftings, it appears that the two preglacial drainage systems became united by an eastward diversion of the western system by the Keewatin ice sheet in the Nebraskan and Kansan stages into the valley of the eastern system. At the culmination of the glacial stages it was largely ice sheet drainage, but later it may have become the course of interglacial drainage from the western area as far down as the lower rapids until the Illinoian glacial stage, in which case its size was similar to that of the present Mississippi.

In the Illinoian stage the valley of the eastern system was covered by the ice sheet below Clinton, Iowa, and also the valley of the western system from Muscatine south to the lower rapids below Fort Madison. So the river was forced to open a new valley outside the border of that ice sheet. This carried the upper Mississippi from above the Illinoian border, as well as the discharge from the melting ice sheet, so its volume somewhat exceeded that of the present river. This was only a temporary course, restricted to the time the valley from Muscatine to Fort Madison was ice-covered. Then, when the Illinoian ice border had shrunk eastward past the Illinois River below Hennepin, the eastern system may have returned to its old course. In that case, the western system would have embraced only the drainage from eastern Iowa and the western edge of Illinois, and would have had a watershed only a small fraction of that of the present Mississippi. This may have been operating throughout the interglacial (Sangamon) stage between the Illinoian and Wisconsin glacial stages, a period of perhaps 100,000 years.

It was in the Wisconsin glacial stage that the Mississippi became permanently established in its present course and drainage area. However, for a few thousand years in the waning part of the Wisconsin stage it received the discharge from the Glacial Lake Agassiz, which apparently more than doubled the flow through the valley.

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"PSEUDOMIGMATITE" IN THE PIEDMONT OF MARYLAND

BY RANDOLPH W. CHAPMAN

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ABSTRACT

The Laurel gneiss lies in the vicinity of Laurel, Maryland, about 17 miles southwest of Baltimore. A cover of Cretaceous gravels obscures all but about 10 square miles. On the north the Laurel gneiss is bounded by Guilford granite and the Wissahickon formation, on the northeast by gabbro, on the east by Cretaceous gravels, on the south by gneiss, and on the west by the Wissahickon formation.

The Laurel gneiss is gray, medium-grained to fine-grained, and faintly foliated. Chief minerals are quartz, oligoclase, biotite, and muscovite. Remnants of schist and quartzite from the Wissahickon formation are common.

Although the Laurel gneiss looks like a true migmatite, it is believed to have originated by the granulation, flowage, and recrystallization of the Wissahickon formation under conditions of stress, high temperature, and abundant water. This is suggested by: (1) the mineralogical and chemical similarity of the two rocks, (2) the granulated and recrystallized texture of the gneiss, (3) the abundance of schist and quartzite remnants which do not show granulation or recrystallization, and (4) the complete gradation from Laurel gneiss into the Wissahickon formation across contacts. The writer thus suggests the name "pseudomigmatite" for this body.

The Laurel gneiss has most of the features of a plutonic igneous rock but it does not show intrusive contacts It is thought, however, that if conditions of metamorphism had been more intense it might have developed even these The origin of the Laurel gneiss appears to be different from that of other igneous-looking rocks.

PROBLEM

In the Spring of 1940 the new geological map of Howard County, Maryland, prepared by Ernst Cloos and Carl Broedel, was published by the Maryland Geological Survey. In the southeast corner of this area is a body of rock which the authors named the "Laurel migmatite".

Cloos believed that it originated by granitization. He pictured the Wissahickon formation as having been thoroughly penetrated and assimilated by intrusions of granitic magma. His interpretation was based mainly upon field studies, however, and, since the "migmatite" has many of the megascopic aspects of an igneous rock with inclusions of schist and quartzite, it is not strange that he should have adopted this view.

Shortly afterward, the present writer undertook a detailed study of the "Laurel migmatite". The areal distribution of the rock was mapped, and the structure and petrography were investigated. As a result, a new theory of origin is advanced. Briefly, the writer believes that the "Laurel migmatite" is not a migmatite but a metamorphic rock which has been produced by the granulation, flowage, and recrystallization of the Wissahickon formation, under conditions of stress, high temperature, and abundant water. The term "migmatite" is, therefore, no longer applicable. However, since the rock has the field appearance of a true migmatite, the writer believes that the term "pseudomigmatite" is more appropriate. In this study the name "Laurel gneiss" is used for this body in preference to a name with more genetic significance.

ACKNOWLEDGMENTS

The writer is especially indebted to Ernst Cloos for his generous assistance in the preparation of this report. He has prepared the section dealing with fabric analyses. Thanks are also due Marland P. Billings and Carleton A. Chapman for reading and criticising the manuscript. Mark H. Secrist, Judson L. Anderson, and James Bell assisted the writer in the preparation of photographs, and this assistance is gratefully acknowledged. George Tunell was kind enough to determine the index of refraction of a specimen of garnet. Through the kindness of Frank L. Hess, spectrographic analyses of two specimens of biotite were made by Morris Slavin at the United States Bureau of Mines. Thin sections for petrographic study were obtained through the generosity of the Maryland Geological Survey.

GEOLOGICAL RELATIONS

The Laurel gneiss lies in the vicinity of Laurel, Maryland, about 17 miles southwest of Baltimore. The body has a length of about 16 miles and extends almost into the city of Washington. The width is uncertain, but it is at least 3 miles. Due to the Cretaceous cover, only about 10 square miles of the gneiss are actually exposed (Pl. 1).

The Laurel gneiss is bounded on the north by the Wissahickon formation and the Guilford "granite" (granodiorite), dikes of which cut the gneiss in many places. To the northeast lies a body of coarse, dark, foliated gabbro whose contact with the Laurel gneiss has never been seen. However, near locality 1 (Pl. 1) the gneiss is cut by a dikelike intrusion which resembles the gabbro petrographically. This suggests that the gabbro is younger. On the east, the Cretaceous sands and gravels of the Coastal Plain lap upon the gneiss and obscure much of its areal extent. The Wissahickon formation bounds the Laurel gneiss on the west and grades into it across a transitional boundary. The structure of these two rocks is closely conformable as may be seen from the geological map (Pl. 1).

A medium-grained, gneissic rock (probably the Sykesville granite of Cloos and Broedel) terminates the Laurel gneiss on the southwest. This gneiss has, locally, some of the features of the Laurel gneiss such as quartzite fragments (inclusions) and quartz lumps, but it differs by being strongly foliated. It is, therefore, probably older than the Laurel gneiss. The boundary between the two gneisses is difficult to locate owing to a lack of good exposures in critical areas and to the similarity of the two rocks. A dashed boundary, located with the aid of outcrops farther to the north, is, therefore, shown on the geological map (Pl. 1).

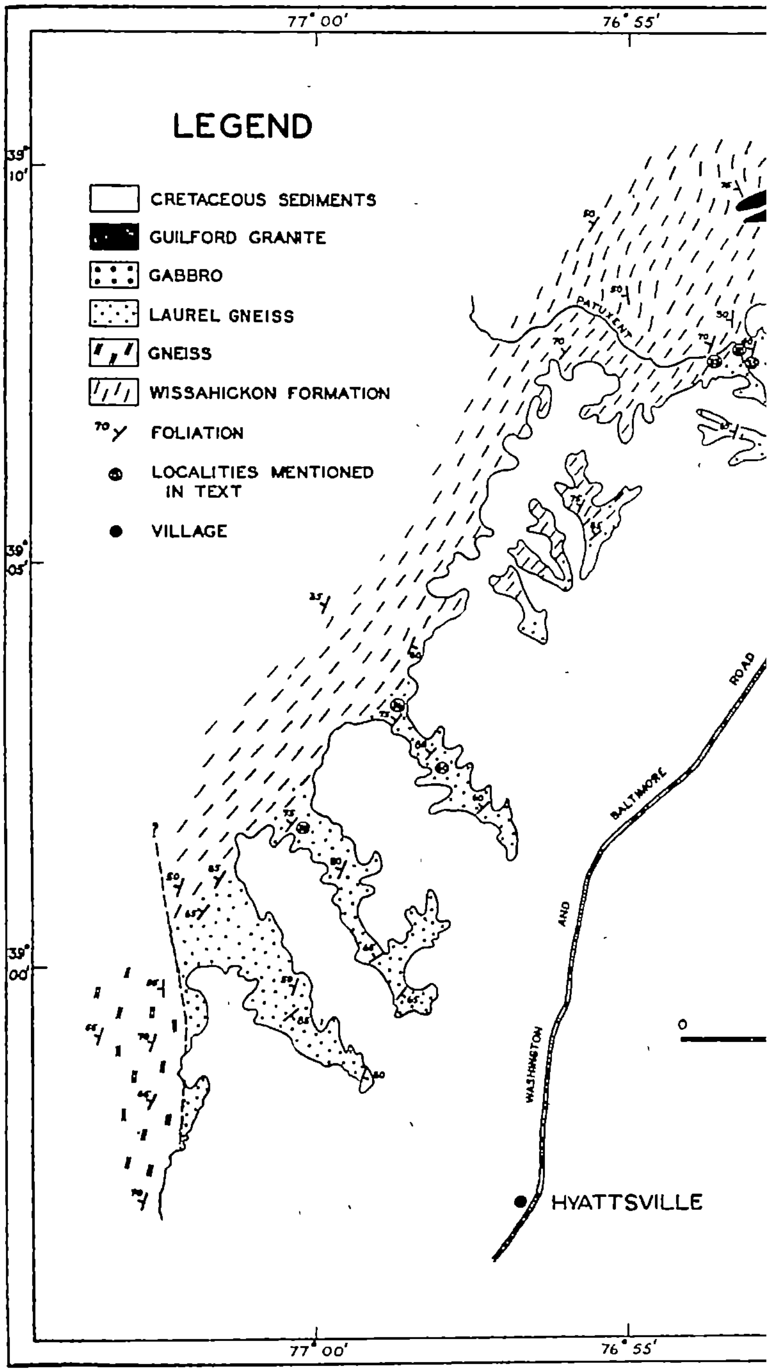
Field localities where specimens were collected, photographs taken, or special features observed are numbered on the geological map (Pl. 1). Since the Laurel gneiss is best exposed in the vicinity of Laurel, most references are made to this locality. However, excellent exposures may also be seen along the two streams flowing southeast from localities 39 and 56.

DESCRIPTION OF THE LAUREL GNEISS

GENERAL FEATURES

The Laurel gneiss is a fine-grained, porphyroblastic rock consisting principally of quartz, plagioclase, muscovite, and biotite. It is light-gray except for those portions which surround mica-rich segregations. (*See* section on segregations.) Where weathered it is yellowish-brown. Megascopically the rock appears fairly uniform in grain size; individual grains range between ¼ millimeter and 1 millimeter in diameter. Some porphyroblasts, however, may reach 2 or 3 millimeters. The texture of the gneiss is granular or sugary.

Structure in the gneiss is not readily apparent everywhere, but many of the fine mica flakes have a rough planar orientation. This produces a foliation which is uniform in strike and dip and conformable with the regional structure. It trends somewhat east of north and dips steeply west. Locally, as for instance at a point 1000 feet northwest of locality 22 (Pl. 1), the foliation is thrown into a series of small but very complex



GEOLOGICAL MAP OF THE LAUREL GNEISS,

folds. In other places it is irregular or wavy. Elsewhere the foliation wraps around remnants of older quartzite or is distorted by them. All these facts suggest strongly that the Laurel gneiss has actually flowed as a plastic mass.

In many places the rock is streaked or banded by long, irregular or lenslike segregations composed principally of coarse biotite and muscovite crystals with diverse orientations. These masses greatly accentuate the foliation of the gneiss since they are streaked out parallel to the latter.

The general appearance of the normal Laurel gneiss in the field is that of an intrusive rock rather than of a metamorphic type. Where remnants of older schist and quartzite are abundant, however, it has more the appearance of a hybrid rock.

MICROSCOPIC FEATURES

Texture —Under the microscope the Laurel gneiss is holocrystalline, inequigranular, and allotriomorphic (Pl 6, fig. 2). Quartz and plagioclase compose much of the rock. The plagioclase crystals are generally equidimensional, but the quartz grains show an elongation parallel to the foliation. Adjacent grains of these minerals are generally interlocked by "sutured" boundaries which are characteristic of recrystallized rocks. Small biotite and muscovite crystals, scattered throughout the groundmass, are roughly oriented, giving the rock its poor foliation. Coarser crystals of biotite and muscovite occur as porphyroblasts cutting the foliation at all angles.

The minerals are distributed unevenly throughout the rock; some areas consist mainly of quartz and plagioclase, and others are very rich in mica. Accordingly, it is impossible to give one mode which will be exactly representative of all specimens. Table 1 shows the modes of several specimens. It will be noted that the same minerals are present in each specimen although they vary in amounts.

Quartz.—Clear, glassy quartz is the most abundant mineral. It occurs as anhedral grains which are elongated parallel to the foliation. The length of the grains (¼ to 1 millimeter) is generally from two to three times the width. Individual crystals are interlocked with surrounding grains of quartz and plagioclase by sutured boundaries. Undulatory extinction is common.

In the transition zone between the Laurel gneiss and the Wissahickon formation large crystals of quartz show special evidence of straining and deformation. Many of these have been broken into fragments. In some instances the fragments have been displaced only slightly. In other cases, however, they are so greatly displaced that evidence of their original relationships is almost lost (Pl. 6, fig. 4).

Plagioclase.—Most of the plagioclase crystals are anhedral and equidimensional. Some appear to be interstitial between quartz grains, suggesting that they crystallized later than the quartz. Most of the plagioclase is calcic oligoclase ($Ab_{72}An_{28}$). One or two sections show a slightly more calcic oligoclase, whereas one contains andesine ($Ab_{63}An_{37}$) Many of the grains are twinned polysynthetically. Straining is much less common than in the quartz. On the whole the oligoclase is clear, but a few grains show alteration to sericite.

Micas —Muscovite and olive-brown biotite are both present in the gneiss although muscovite is somewhat more abundant. There are two generations of each mica.

The small crystals of the younger generation, scattered more or less evenly throughout the rock groundmass, are clear and free from inclusions. They have a rough planar orientation. Large porphyroblasts of the later generation show diverse orientations. The biotite porphyroblasts have frayed edges and contain abundant poikiloblastic inclusions of quartz, plagioclase, magnetite, and zircon. In some, inclusions are so abundant that the porphyroblasts are mere skeletons. The muscovite porphyroblasts have more regular boundaries than the biotite porphyroblasts. They contain inclusions of quartz, plagioclase, garnet, magnetite, apatite, and zircon. Many muscovite and biotite porphyroblasts are intergrown with one another parallel to cleavage faces suggesting simultaneous crystallization. Both the muscovite and biotite porphyroblasts show wavy extinction.

The optical properties of the muscovite are as follows: $\beta = 1.605$, $\gamma = 1.610$; optically negative; $2V = 40°$; dispersion $r < v$, weak. These properties are similar to those given by Barth (1936, p. 780-781) for a muscovite from Dutchess County, New York, and suggest that the chemical composition is approximately the same. Barth's analysis of the Dutchess County muscovite is as follows:

SiO_2	46.31
TiO_2	0.79
Al_2O_3	38.52
Fe_2O_3	tr
FeO	.
MgO	0.18
CaO	0.46
Na_2O	1.56
K_2O	8.03
Li_2O	tr
H_2O	4.46
Total	100.31

Since the optical properties of the muscovite porphyroblasts are the same as those of the oriented muscovite crystals in the groundmass, their chemical composition is doubtless the same.

Optical study of the biotite shows the following properties: $\alpha = 1.590$, $\beta = 1.641$, $\gamma = 1.641$; optically negative; $2V = $ ca. 3°; dispersion $r < v$, medium to strong; pleochroism, X = pale yellow-brown, Y = olive-brown, Z = olive-brown, $X < Y = Z$. However, the large porphyroblastic biotites which have been strained have an optic angle ranging from 10° to 15°. According to Winchell (1933, p. 274, Fig. 200) the above optical properties indicate common biotite with FeO:MgO = 55:45. Both the biotite porphyroblasts and the fine, oriented biotites have the same optical properties so their chemical compositions are probably the same. A few slides show biotite altering to chlorite.

Garnet.—Pink garnet occurs chiefly as large, rounded crystals. In many slides the fine, oriented micas of the groundmass were seen to bend around the garnet crystals. proving that in general the garnets are older. In one or two instances, however, garnets appeared to cut across quartz, plagioclase, and mica. Poikiloblastic inclusions of quartz, magnetite, biotite, and muscovite have been found in the garnet. The index of refraction is 1.815± .010. According to Eskola's diagram (1921, p. 9) this indicates almandine with the following composition:

Mol. per cent	
87	$Fe_3Al_2Si_3O_{12}$
5	$Mg_3Al_2Si_3O_{12}$
8	$Ca_3Al_2Si_3O_{12}$

Tourmaline.—Tourmaline is an accessory in the Laurel gneiss, but optical data were obtained in order that this mineral could be compared with that in the Wissahickon schist. These optical properties are as follows: $\varepsilon = 1.631$ (pale yellow-pink), $\omega = 1.661$ (deep blue-green). Larsen and Berman (1934, p. 247) describe a tourmaline with similar optical properties which has the following chemical composition.

$$(Na, Ca)\ R_3\ Al_6\ B_3\ Si_6\ O_{27}\ (OH)_4$$

in which Na.Ca = 12:1, and R = Mg:Fe″:Fe‴ = 2:3:5.

Accessory minerals.—In addition to the tourmaline, accessory minerals consist of magnetite, zircon, apatite, chlorite, and sericite.

REMNANTS

Introduction.—The Laurel gneiss contains many remnants of the Wissahickon formation. The reason for calling these remnants rather than inclusions will be realized later. They are of three different types: (1) coarse biotite-muscovite schist, (2) fine biotite-muscovite quartzite, and (3) quartz lumps.

Schist.—The remnants of coarse biotite-muscovite schist may be studied best at locality 22 and along the Patuxent River to the northwest (Pl. 1). They occur as bands or irregular masses with ragged boundaries and range from a few inches up to 10 feet or more in width. Most of the remnants are oriented with their long axes parallel to the foliation. The number of remnants differs greatly. In some of the outcrops northwest of locality 22 (Pl. 1) remnants comprise about 50 per cent of the rock. In other outcrops in this vicinity the remnants are few and scattered. Some of the exposures around locality 22 (Pl. 1) are so rich in remnants that the rock appears more like the Wissahickon schist than it does like the Laurel gneiss.

The schist remnants are coarse, recrystallized, porphyroblastic, and distinctly foliated. The minerals of the groundmass have an average diameter of about 1 millimeter, but because of abundant porphyroblasts the rock appears much coarser. The porphyroblasts consist chiefly of large muscovite crystals with diameters of from 5 to 15 millimeters which give the remnants a brilliant spangled appearance. Individual crystals of tourmaline are conspicuous, some attaining a length of nearly 2 centimeters. Other porphyroblasts are red garnets (diameter 1 to 2 millimeters) and biotite (diameter 2 to 3 millimeters).

Under the microscope the schist remnants are inequigranular and decidedly schistose. They consist of muscovite-rich layers which alternate with layers of quartz and calcic oligoclase. The muscovite crystals in the groundmass show an almost perfect planar orientation. The coarse porphyroblasts of muscovite, garnet, olive-brown biotite, and blue-green tourmaline cut directly across the structure of the groundmass indicating clearly that they are younger.

There can be little doubt that these remnants represent fragments of Wissahickon schist. In the first place, the textures, modes, and sequences of crystallization of the two rock types are the same. Furthermore, optical data suggest that the minerals of the remnants have the same chemical composition as those of the Wissahickon schist.

Quartzite.—Light gray, fine-grained quartzite remnants also are abundant in the Laurel gneiss. These may be observed best at localities 20, 22, and 35 (Pl. 1). Many are elongate with a length of from 1 inch to 30 inches and a width of approximately one-third their length. Others are more equidimensional or irregular with diameters of from 1 to 12 inches. Most of the quartzite remnants show fine banding. In those which are elongate the banding is generally parallel to their longest dimension, which in turn is almost always parallel to the foliation of the gneiss. Many of the equidimensional or irregular ones, however, lie with the fine banding of the quartzite at an angle to the foliation. Several of these may be seen at locality 35 (Pl. 1). It is important to note that near the remnants the foliation in the Laurel gneiss commonly bends around parallel to the boundaries. This indicates clearly that the foliation is younger than the remnants.

The contact relations between a fine-grained quartzite remnant and the surrounding Laurel gneiss are also shown at locality 11 (Fig. 1). Here an irregular quartzite block, 6 inches across, and with no visible foliation, is inclosed in the gneiss. For the most part, the foliation of the gneiss is parallel to the boundaries of the block. Near the lower end of the quartzite block, the foliation of the gneiss has been thrown into a series of small folds. It is clear, therefore, that the foliation of the Laurel gneiss is younger than the quartzite remnant since it has been deformed by the latter.

One quartzite remnant at locality 35 (Pl. 2, fig. 2) has been faulted and displaced about 4 inches. The foliation of the Laurel gneiss, which wraps around the remnant, has also been displaced, but the break has been sealed by recrystallization.

The quartzite composing the remnants is a light-gray, sugary rock with a grain size of from ⅛ to ½ millimeter. Microscopically it is equigranular consisting mainly of interlocking grains of strained quartz with some plagioclase. Minute crystals of olive-brown biotite and somewhat larger grains of muscovite show a subparallel orientation. A few large muscovite, biotite, and garnet crystals form porphyroblasts.

The remnants described above are similar both texturally and mineralogically to the fine-grained quartzite layers interbedded with the Wissahickon schist. Modes of two of these quartzite layers (20B and 25) are shown in Table 1. This similarity leaves little doubt the fine-grained masses in the Laurel gneiss are remnants of Wissahickon quartzite.

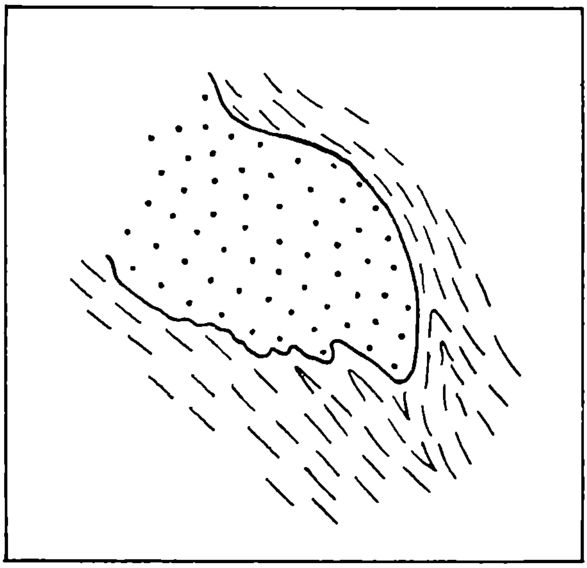


FIGURE 1.—*Quartzite remnant in Laurel gneiss*

Sketch, drawn at locality 11 (Pl 1), showing the effect of a Wissahickon quartzite remnant (dotted) on the foliation of the gneiss (dashed). Width of quartzite remnant about 6 inches

Quartz lumps.—Small irregular lumps of glassy or milky quartz may be studied best at locality 20 and along the Patuxent River to the west (Pl. 1). Their diameter ranges from a quarter of an inch to 4 inches; the majority have a diameter of about 1 inch. Some of the quartz lumps, are elongate or lenslike with the longer axis parallel to the foliation. Others, however, are spheroidal or irregular and show no special orientation. The lumps stand out on weathered surfaces because they are more resistant than the surrounding gneiss.

Under the microscope each lump is seen to consist of a group of large, highly strained quartz grains (Pl. 6, fig. 3) completely interlocked with one another by sutured boundaries, thus indicating thorough recrystallization. The boundary of the lump is likewise sutured as a result of intercrystallization of the lump with the crystals of the surrounding gneiss

The quartz lumps are most likely recrystallized remnants of thin quartz layers and veins which occur in the Wissahickon formation.

SEGREGATIONS

Dark segregations are common in the Laurel gneiss. These consist of accumulations of large porphyroblasts of biotite and muscovite with occasional garnets. Individual crystals vary from 1 to 5 millimeters in diameter. In general, they show random orientations.

The segregations are of two different types. Most common are mica masses pulled out into dark, irregular streaks or lenses more or less parallel to the foliation of the gneiss. These streaks have an average width of from 1/8 to 1 inch, and a length of from 1 inch to more than 1 foot. Where the streaks are abundant the gneiss is strongly banded.

The second type of segregation is best demonstrated at locality 20 and at a point on the Patuxent River 500 feet west (Pl. 1). At these localities the masses of unoriented mica are rounded or irregular, and their diameter varies from 1/4 inch to several inches. The dark segregations are surrounded by a zone of cream-colored material, 1/2 inch to 4 inches wide (Pl. 4, figs. 1, 2; Pl. 5, figs. 1, 2). Thin sections show that this light-colored zone differs from the normal gray gneiss only because it is relatively deficient in both biotite and muscovite. It is quite evident, therefore, that this type of segregation has formed at the expense of the micas in the immediately surrounding gneiss. Its origin will be considered later.

WISSAHICKON FORMATION

Field and laboratory studies reveal a close genetic relationship between Laurel gneiss and the Wissahickon formation. In order to demonstrate this relationship it will be necessary first to describe the latter briefly.

The Wissahickon formation which bounds the Laurel gneiss on the west is an oligoclase-mica schist. It consists chiefly of strongly folded mica schist interbedded with thin layers of fine-grained micaceous quartzite. A strong schistosity is characteristic of the mica schist. The quartzite, however, is only faintly banded by biotite-rich layers. These structures are regional and trend slightly east of north and dip steeply west (Pl. 1).

The mica schist is a completely recrystallized, coarse- to medium-grained rock which consists chiefly of muscovite, biotite, quartz, and calcic oligoclase ($Ab_{70}An_{30}$). Its grain size varies from 1 millimeter to several millimeters. Mica-rich layers (1/2 millimeter thick) alternate with layers rich in quartz and oligoclase (1/2 millimeter to 2 millimeters thick). The micas are in large flakes which are essentially oriented in a plane, but which locally wrap around large garnet crystals or lenticular areas of quartz. Practically all the quartz grains show strain

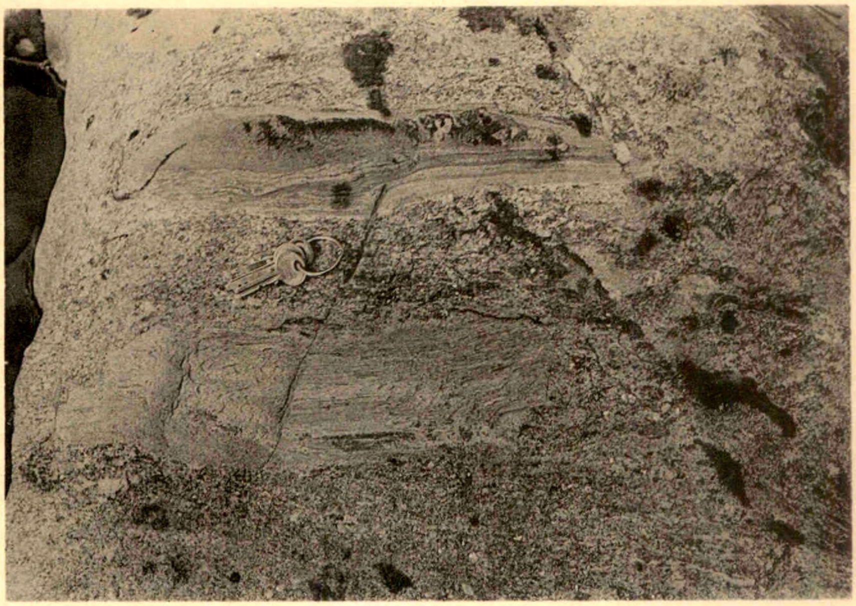


FIGURE 1. LAUREL GNEISS WITH REMNANTS OF FINE WISSAHICKON QUARTZITE
Remnants parallel to crude foliation of gneiss. Note that remnants look like inclusions in an igneous rock. Outcrop in Patuxent River 350 feet south of locality 35 (Pl. 1).

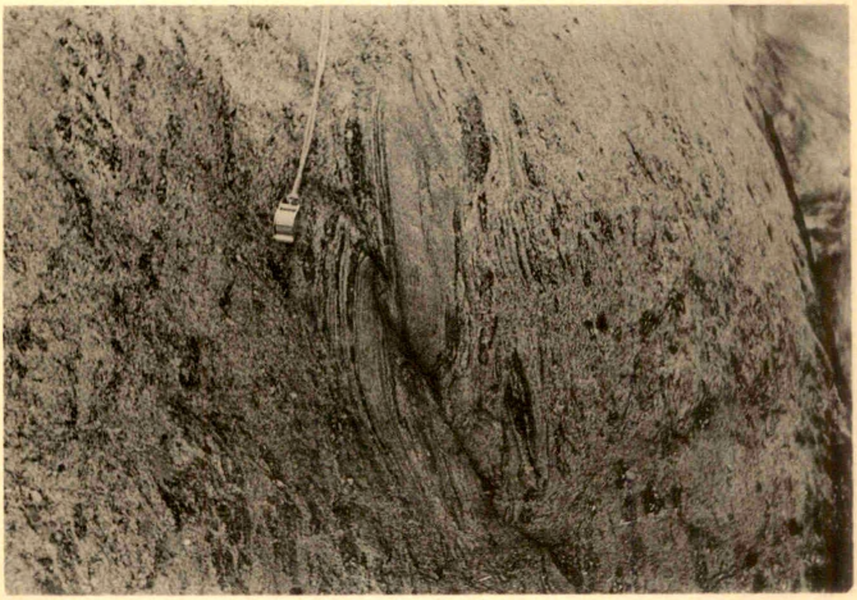


FIGURE 2. FAULTED QUARTZITE REMNANT WITH FOLIATION OF LAUREL GNEISS WRAPPING ABOUT IT
Note how the fault plane near the hand lens has been sealed over by later flowage and recrystallization. Appearance of gneiss is similar to that of an igneous rock. Outcrop at locality 35 (Pl. 1).

LAUREL GNEISS WITH REMNANTS OF WISSAHICKON QUARTZITE

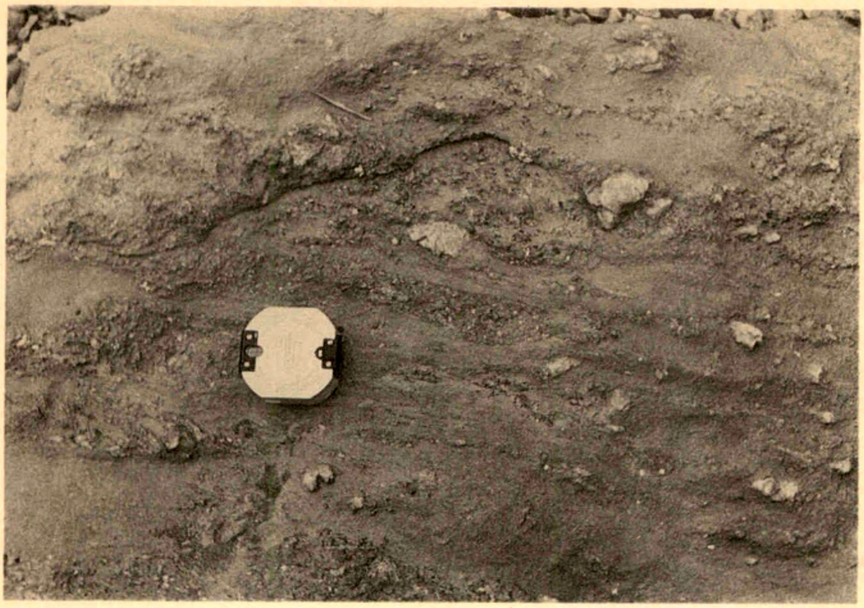


FIGURE 1. FINE-GRAINED LAUREL GNEISS WITH REMNANTS OF COARSE WISSAHICKON SCHIST
White, irregular quartz lumps are also remnants from Wissahickon formation. Picture suggests a hybrid rock produced by fine, even-grained granite penetrating coarse schist. Outcrop in Patuxent River 1000 feet northwest of locality 22 (Pl. 1).

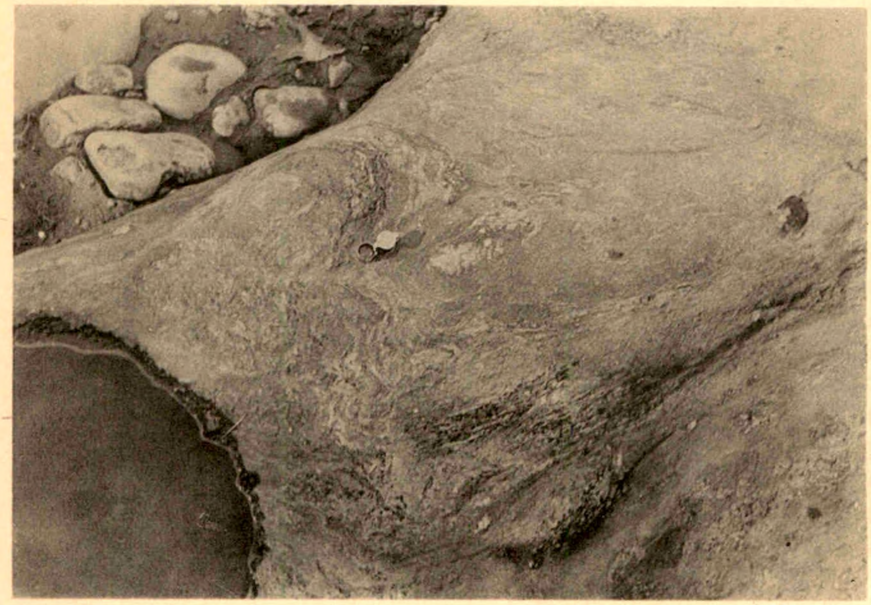


FIGURE 2. LAUREL GNEISS WITH CONTORTED FLOW BANDS
Outcrop in Patuxent River 1000 feet northwest of locality 22 (Pl. 1).

LAUREL GNEISS WITH WISSAHICKON SCHIST REMNANTS, QUARTZ LUMPS, AND FLOW BANDS shadows. Large porphyroblasts of muscovite, biotite, garnet, and tourmaline are common and show random orientations. They are filled with abundant inclusions of finer minerals of the groundmass. Accessory minerals of the schist include zircon, apatite, magnetite, and pyrite. Staurolite and cyanite have been found in this phase of the Wissahickon formation in some places, but they have not been observed close to the Laurel gneiss. Modes of the schist are shown in Table 1. It will be noted that the mineral percentages vary somewhat, although the same kinds of minerals are present in all specimens.

The quartzite phase of the Wissahickon formation is a light-gray, equigranular, faintly banded rock. Its grain size varies from 1/8 to 1/2 millimeter. The rock consists essentially of quartz and oligoclase with suboriented plates of muscovite and olive-brown biotite. Porphyroblasts of muscovite, biotite, and garnet are present in some specimens. Accessory minerals include zircon, apatite, magnetite, garnet, epidote, and chlorite. Modes of the quartzite are shown in Table 1.

RELATION OF THE LAUREL GNEISS TO THE WISSAHICKON FORMATION

MINERAL COMPOSITION

Table 1 shows the modes of three specimens of Laurel gneiss and several specimens of schist and quartzite from the Wissahickon formation near by. Five points regarding these should be emphasized.

(1) The three specimens of gneiss (5, 9, and 12A) show essentially the same minerals although in different proportions. It will be noted that the mineral assemblage in these specimens is not that of any typical igneous rock.

(2) The modes of the Wissahickon formation are variable in mineral proportions, but this is to be expected since they represent a sedimentary series which consists of alternating layers of schist and quartzite. Quartzite layers are not numerous in the Wissahickon formation, however, and are always thin. Detailed studies indicate that in the vicinity of Laurel, Maryland, the Wissahickon formation consists of approximately 90 per cent schist and 10 per cent quartzite. Specimens 1 and 11 were chosen because they are believed to be most representative of the schist phase of this formation. Specimens 20B and 25 were chosen as typical samples of the quartzite phase. The two schist specimens show a close modal similarity, but the two quartzite specimens differ somewhat in their mineral proportions. The quartzite specimens are richer in quartz and poorer in mica than the schist.

(3) There is an unquestionable similarity in certain respects between the modes of the Laurel gneiss (5, 9, and 12A) and those of the Wissa-

hickon formation. Exactly the same minerals occur in the gneiss that are present in the schist and quartzite. Furthermore, in the two samples of typical schist (1 and 11) even the relative proportions of these minerals are roughly the same as in the modes of the gneiss. However, the quartzite specimens (20B and 25) differ considerably from the gneiss in mineral ratios. These relations are exactly as would be expected if, as the writer believes, the Laurel gneiss has been derived from the Wissahickon formation. By granulation, flowage, and recrystallization the heterogeneous Wissahickon formation has been thoroughly mixed to produce a more or less homogeneous rock. Since the quartzite layers are not numerous (about 10 per cent of the Wissahickon formation), they would not make the gneiss much richer in silica than the typical schist.

Table 1.—*Modes of Laurel gneiss, schist, and quartzite from Wissahickon formation, and Guilford granite**

(Volume per cent)

	5	9	12A	1	11	16	20B	25	8
Quartz	40.9	48.2	37.7	43.8	40.9	45.2	66.6	65.0	30.2
Plagioclase	16.3	22.2	21.5	19.2	20.1	13.1	9.4	16.5	33.5
	An_{32}	An_{28}	An_{30}	An_{28}	An_{28}	An_{27}	An_{28}	An_{28}	An_{10-15}
Biotite	16.6	11.2	11.5	12.3	14.0	15.2	4.8	8.4	3.5
Muscovite	24.1	16.9	28.2	22.4	23.0	22.4	15.7	7.8	tr
Garnet	0.7	0.7	....	1.6	1.5	3.7	3.3	....	....
Magnetite	1.4	1.0	1.1	0.7	0.5	0.4	0.2	1.8	tr
Tourmaline	....	....	....	tr	....	tr	....	....	....
Zircon	tr	tr	tr	tr	tr	tr	tr	tr	tr
Apatite	tr	tr	tr	tr	tr	tr	....	tr	tr
Chlorite	tr	tr	tr	tr	tr	tr	tr	....	....
Epidote	....	....	....	....	....	....	....	0.5	....
Microcline	....	....	....	....	....	....	....	....	32.8

* All modes measured with a Wentworth integrating stage.

All locality references are to Plate 1.

5 Laurel gneiss, 800 feet west-southwest of locality 1.
9 Laurel gneiss, locality 9.
12A Laurel gneiss, locality 20.
1 Schist from Wissahickon formation, locality 1.
11 Schist from Wissahickon formation, 1/3 mile west-southwest of locality 9.
16 Schist-gneiss, locality 28. Transition product between Laurel gneiss and schist of Wissahickon formation.
20B Quartzite from Wissahickon formation, locality 33.
25 Quartzite from Wissahickon formation, near locality 33.
8 Guilford granite, locality 8.

(4) Specimen 16 is from the transition zone between the Laurel gneiss and the Wissahickon formation. It is believed to represent schist partly transformed to gneiss. Its mode is similar to those of both these rock types.

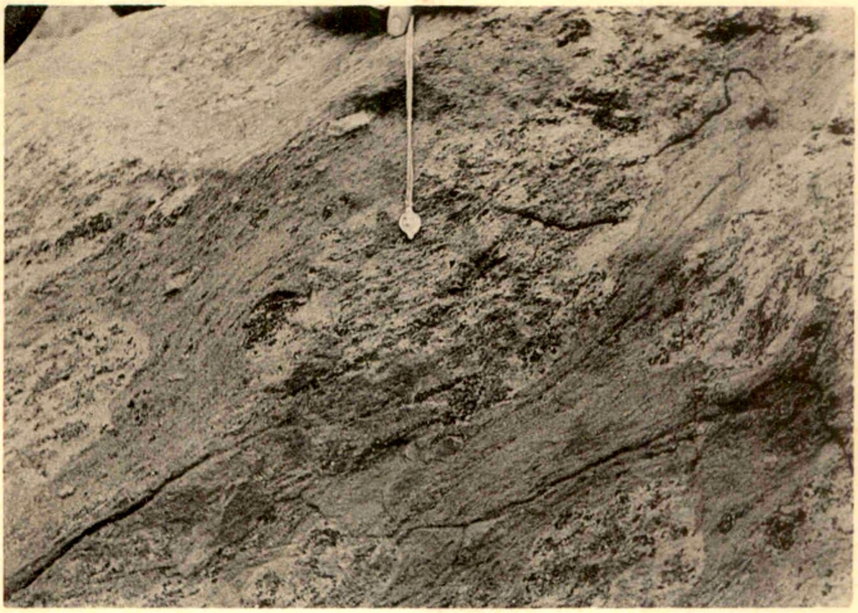


FIGURE 1. FINE-GRAINED, GRAY, FOLIATED LAUREL GNEISS
Enclosing elliptical areas which consist of mica segregations surrounded by cream-colored gneiss. Outcrop in Patuxent River 500 feet west of locality 20 (Pl. 1).

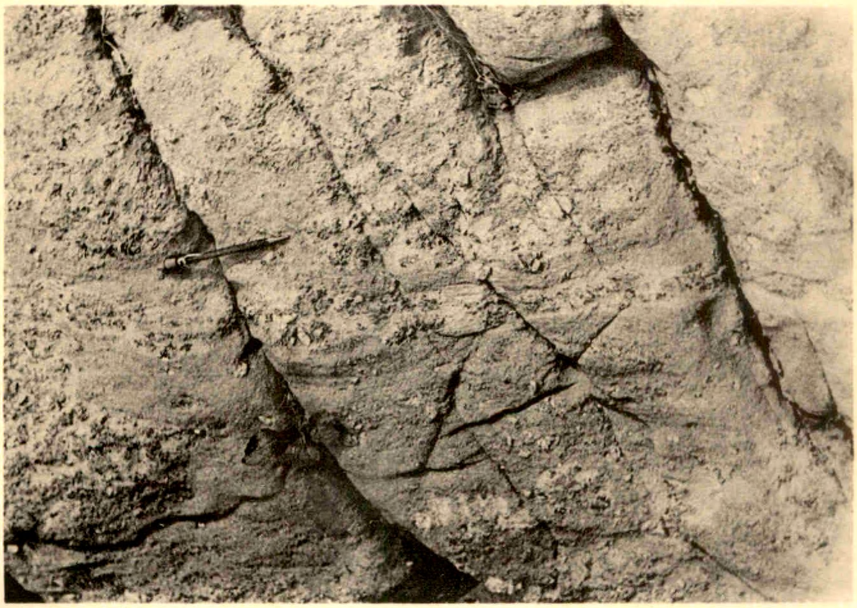


FIGURE 2. FINE-GRAINED, GRAY LAUREL GNEISS
With areas of light-colored gneiss containing mica segregations. Segregations have formed at the expense of the micas in the normal gray gneiss, thus producing light areas. Outcrop in Patuxent River 500 feet west of locality 20 (Pl. 1).

MASSES OF SEGREGATIONS IN THE LAUREL GNEISS

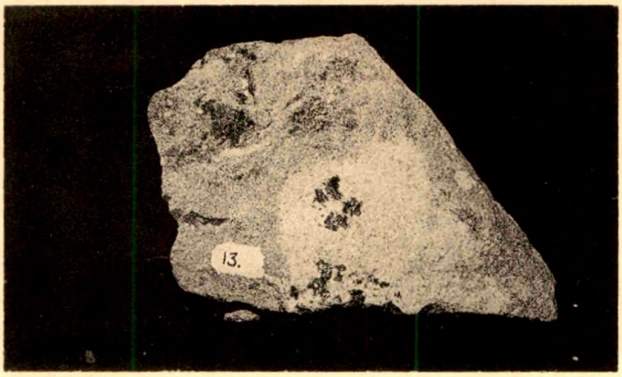


Figure 1. Specimen of Laurel Gneiss (gray)
Containing mica segregations (black spots) surrounded by cream-colored zone. Segregations consist of biotite and muscovite porphyroblasts accumulated by aqueous solutions. Material of the segregations was derived from groundmass micas of gneiss, thus leaving a light-colored zone around each segregation. Height of specimen is 3½ inches. From outcrop in Patuxent River 500 feet west of locality 20 (Pl. 1).

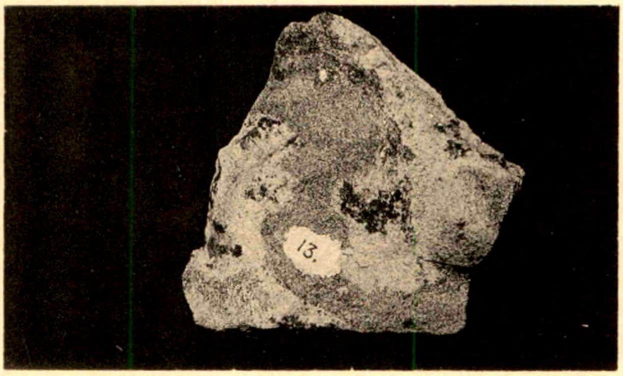


Figure 2. Same as Figure 1 of Plate 5
Height of specimen is 4 inches.

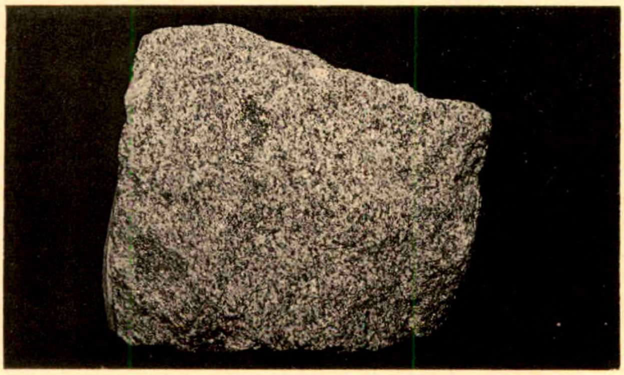


Figure 3. Specimen of Typical Laurel Gneiss
Note apparent similarity to a plutonic igneous rock. Height of specimen is 3½ inches. From locality 20 (Pl. 1).

MICA SEGREGATIONS AND TYPICAL LAUREL GNEISS

(5) The optical properties of the plagioclase, muscovite, biotite, garnet, and tourmaline in the Wissahickon formation have been determined accurately. These optical properties are identical with those of the corresponding minerals in the gneiss. This indicates that the minerals of the two rocks are similar in chemical composition.

Spectrographic determinations were made of the alkali content of the biotite from both the Laurel gneiss and the schist from the Wissahickon formation. Biotite was separated from specimen 12A of the gneiss from locality 20, and from a specimen of coarse schist from an outcrop 7 miles northwest of the northern end of the gneiss body (1 mile northwest of the village of Clarksville). The alkali content was found to be the same in both biotites within the range of spectrographic methods. The results are as follows:

Material	Approximate percentage
K_2O	5
Rb_2O	1
Li_2O	1
Na_2O	3
Cs_2O	not detected

This evidence corroborates the optical data in indicating that the biotite in the two rocks is the same.

CHEMICAL COMPOSITION

In comparing the chemical composition of the Laurel gneiss with that of the Wissahickon formation it was found impracticable to make chemical analyses of the rocks. From rocks which are as variable as these two, it would be impossible to select a few specimens for analyses which would be thoroughly representative of each type. Many chemical analyses of well-chosen specimens would be necessary in order to determine definitely any chemical relationship between the Laurel gneiss and the Wissahickon formation. Such chemical analyses would involve an expense all out of proportion to the benefits derived.

The author has thought of using composite samples, each consisting of approximately 20 pieces of equal weight, of the two principal rock types: Laurel gneiss and Wissahickon schist. However, this possibility was discussed with a number of leading petrographers, and they all recognized the problem of proper sampling. Plates 2, 3, and 4 will readily convince anyone of the difficulties of sampling in rocks of which no 2 cubic feet are alike. Inasmuch as all the minerals in both rocks are exactly of the same composition and there are none in one rock that do not occur in the other, chemical analyses could at best show the differences in proportion of the otherwise identical minerals. There is little doubt in the author's mind that analyses would show nothing but would merely indicate the chemical similarity of these two rock types.

For these reasons it was considered more practicable to calculate the average chemical composition of the Wissahickon formation and the Laurel gneiss from a study of several thin sections. Since the mineralogy of both rock types is relatively simple, the results obtained are believed to be fairly accurate.

TABLE 2.—*Chemical compositions*

	A	B	C
SiO_2	72.87	71.10	70.18
TiO_2	......	......	0.39
Al_2O_3	16.27	17.10	14.47
Fe_2O_3	0.48	0.76	1.57
FeO	1.54	1.72	1.78
MnO	......	......	0.12
MgO	1.08	1.13	0.88
CaO	1.07	1.21	1.99
Na_2O	1.61	1.65	3.48
K_2O	3.95	4.29	4.11
P_2O_5	......	......	0.19
H_2O	1.13	1.04	0.84
	100.00	100.00	100.00

A Average chemical composition of the Wissahickon formation near Laurel, Maryland. Calculated from modes of schist and quartzite shown in Table 1.

B Average chemical composition of Laurel gneiss. Calculated from the three modes shown in Table 1.

C Chemical composition of average granite of all periods (Daly, 1933, p. 9).

After a study of several thin sections, specimens 1 and 11 (Table 1) were chosen as most representative of the schist phase of the Wissahickon formation, and specimens 20B and 25 were chosen as representative of the quartzite phase. It was believed better to choose representative thin sections rather than to average all of them since some sections were prepared from unusual rock types. The modes of specimens 1 and 11 were then averaged to give the average mode of the schist. Likewise, the average mode of quartzite was obtained by averaging the modes of specimens 20B and 25. Now it has been pointed out that the schist makes up about 90 per cent of the Wissahickon formation, and the quartzite about 10 per cent. Thus by taking the average mode of the schist and the average mode of the quartzite, in a ratio of 9 to 1, the average mode of the whole Wissahickon formation was obtained. From this average mode, the average chemical composition of the Wissahickon formation was calculated (Table 2, column A).

The average chemical composition of the Laurel gneiss was determined in a similar way. Several thin sections of the rock were studied, and specimens 5, 9, and 12A (Table 1) were chosen as representative of the whole body. The modes of these three specimens were then averaged so that the average mode of the gneiss was obtained. From this mode, the average chemical composition of the Laurel gneiss was calculated (Table 2, column B).

A study of Table 2 shows a surprisingly close similarity between the calculated average chemical composition of the Laurel gneiss (column B) and that of the Wissahickon formation (column A). The author's contention that the Laurel gneiss was produced by the thorough mixing and recrystallization of the various rock types of the Wissahickon formation is fully in line with this evidence.

Column C in Table 2 gives the chemical composition of the average granite of all periods as calculated by Daly. It will be noted that the composition of the Laurel gneiss (column B) is similar to that of the average granite except that alumina is somewhat higher and soda somewhat lower in the former. The significance of this point will be noted later.

STRUCTURAL AND AGE RELATIONS

The foliation of the Laurel gneiss is essentially parallel to that of the Wissahickon formation (Pl. 1). The following observations indicate, however, that the gneiss and its foliation are younger than the Wissahickon formation and its foliation. In the first place, the schist of the Wissahickon formation is a coarse, thoroughly recrystallized, mica-rich rock with a well-developed schistosity or foliation. This foliation is secondary and has been produced by intense regional metamorphism. The Laurel gneiss, on the other hand, does not show this marked schistosity so it must be younger. It is finer-grained and shows only a vague foliation which has more the appearance of a primary flow structure. In the second place, remnants of quartzite and strongly foliated schist from the Wissahickon formation are common in the gneiss. Many of these are oriented so that their foliation is parallel to the poorer foliation of the gneiss; others are unoriented. Finally, at localities 11 and 20 (Pl. 1) the foliation of the gneiss bends around remnants of quartzite which are themselves banded. (*See* discussion of quartzite remnants.) The foliation of the Laurel gneiss has clearly been deformed around the older banded remnants.

FABRIC ANALYSES

BY ERNST CLOOS

Introduction.—Fabric studies of the Laurel gneiss and Wissahickon schist indicate three significant things. First, the mica orientation in the schist and in the gneiss are essentially the same. This suggests that the directions of movement which produced mica orientation in the gneiss are nearly the same as those which caused orientation in the schist. Secondly, the c axes of quartz in the schist are essentially perpendicular to the foliation of the schist. In the gneiss, however, the largest concentration of quartz axes lies in the foliation planes produced by the micas, whereas a second group is oriented at an angle of about 45 degrees to these planes. It appears that the quartz in the gneiss represents original large grains which have been fractured along crystal directions and oriented according to shape by rock flowage. This supports the concept that the Laurel gneiss represents the Wissahickon formation after granulation, plastic deformation, and later recrystallization. Finally, the fabric of the schist is definitely that of a metamorphic rock (B-tectonite), whereas the quartz orientation in the Laurel gneiss is similar to that in an igneous rock.

Method.—Biotite, muscovite, and quartz orientations were measured with the universal stage in the conventional manner. The lower half of an equal-area projection was used, and counting at each half-centimeter intersection was found more accurate than the usual method of counting at each centimeter intersection. The preparation and interpretation of fabric diagrams is adequately described in the literature by Haff (1938), Knopf and Ingerson (1938), and Fairbairn (1939), and a detailed account of the method seems superfluous here.

One specimen each of schist from the Wissahickon formation (locality 33, Pl. 1) and the Laurel gneiss (450 feet northeast of locality 33, Pl. 1) were investigated. Thin sections were cut as closely as possible at an angle of 90 degrees to the strike of the megascopic foliation. Strike and dip of the projection planes are marked with an arrow and a short line in the diagram. East and west are marked E and W respectively. All strike readings were measured clockwise from N (= 0°) to S (= 180°). For example: N 45°E = 45°, and N 45°W = 135°, with the arrow pointing northeast and northwest respectively.

The mica diagrams include 100 grains each, and since the orientation is very pronounced this number is thought to be sufficient in ascertaining the principal orientation of cleavage poles. In the schist 400 average-sized quartz grains were measured, and 200 large quartz grains were measured in the gneiss.

Figure 5 shows the orientation of 100 quartz axes in relation to the shape of the quartz grains in the gneiss. Since many of the grains are elongated, the dimensions were measured with a micrometer, and the ratio between the longest and shortest form axes was plotted in the location of the optical axes. In this manner the shape of each grain appears in relation to its optical orientation. The diagram is not a counted diagram.

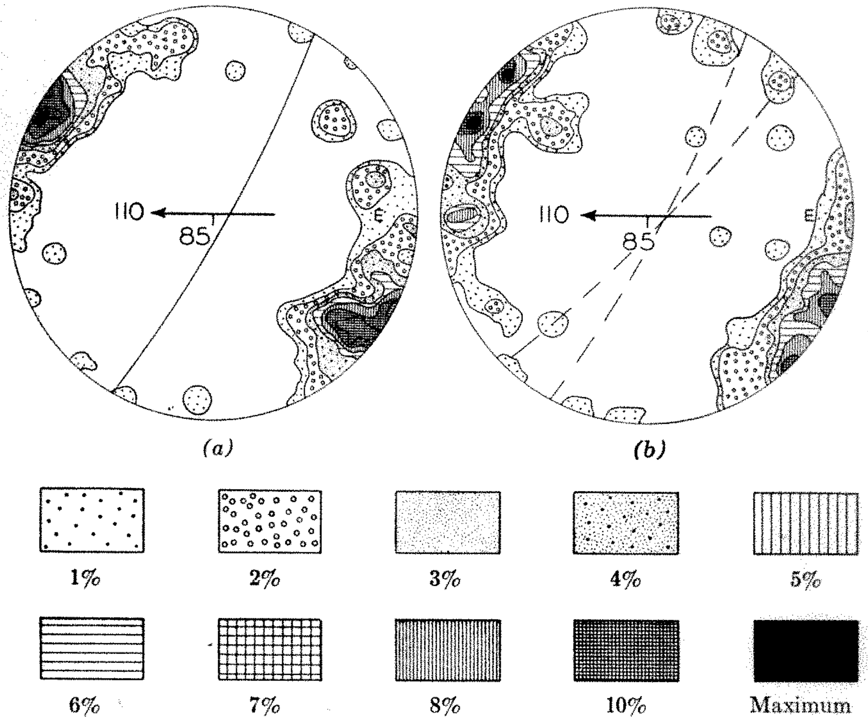


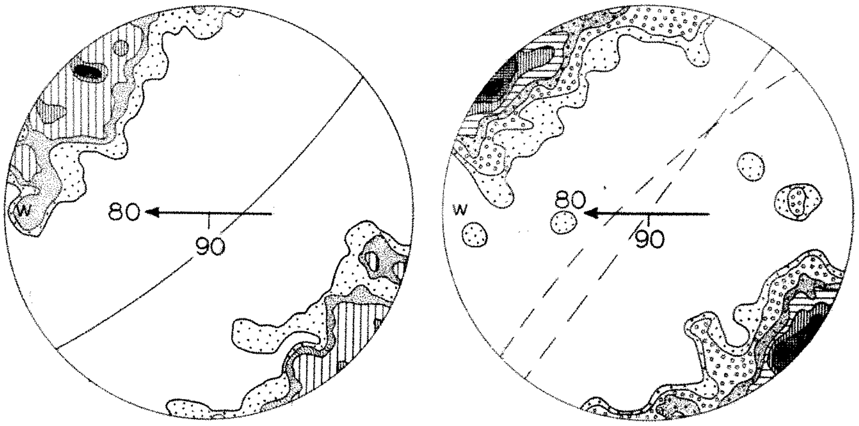
FIGURE 2.—*Muscovite and biotite diagrams of Wissahickon schist*

(*a*) Muscovite from Locality 33 (Pl. 1). 100 muscovite cleavage poles, excluding porphyroblasts. Contours 1-2-4-6-8-10-12(15). Trace of cleavage plane indicated by solid line.

(*b*) Biotite from Locality 33. 100 biotite cleavage poles. Contours 1-2-4-6-8-10-12(13). Traces of biotite cleavages indicated by broken lines. Girdle interrupted, axis of rotation about 10° E. of center.

Mica orientation.—All micas measured are those in the groundmass; no porphyroblasts were included. The poles of biotite cleavages in both rocks are concentrated in two distinct maxima, representing two planes which intersect at angles of 10 and 15 degrees (Figs. 2b, 3b). A girdle is suggested but is incomplete. In spite of the small angle between the planes, and the high concentration around the maxima, the writer believes that the concentration of biotite in two planes is significant since this freedom of orientation has been found also in more than 100 other localities throughout the Wissahickon formation.

The muscovite cleavage poles in both rocks are concentrated in only one maximum (Figs. 2a, 3a) which does not necessarily coincide with any of the biotite maxima. In other areas of the Wissahickon formation large deviations up to 30 degrees have been measured, and these are thought to be due to a rotation of the fabric between stages of recrystallization. In these areas muscovite is later than biotite, and its orientation is always more distinct (Cloos and Hietanen, 1941).



(a) (b)

FIGURE 3.—*Muscovite and biotite diagrams of Laurel gneiss*

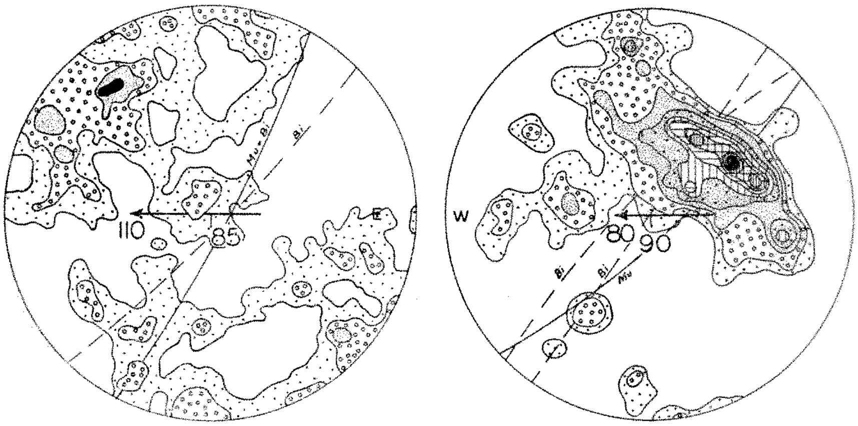
(*a*) Muscovite from 450 feet northeast of locality 33 (Pl. 1). 100 muscovite cleavage poles. Contours 1-3-5-8-10. Trace of cleavage indicated by solid line.

(*b*) Biotite from same locality as 3a. 100 biotite cleavage poles. Contours 1-2-4-6-8-10(12-16). Broken lines indicate traces of biotite cleavages intersecting in quartz maxima of Figure 4b.

See Figure 2 for contour shading.

In the present investigation no age difference between biotite and muscovite was noted, and this may explain the unusually small deviation between orientations. As a whole, biotite is not so well oriented as muscovite. The arrangement of mica poles in both schist and gneiss represents a well-defined megascopic schistosity and foliation respectively (Figs. 2a-3b).

Quartz orientation.—The orientation of quartz grains differs widely in the two rock types (Figs. 4a, 4b). In the schist (Fig. 4a) 400 average-sized grains are concentrated in an incomplete girdle which covers half the circumference and culminates in a 4 per cent maximum, near the location of the cleavage poles of the micas. This shows preferred orientation of quartz grains roughly perpendicular to the cleavage planes. This orientation is described by Fairbairn (1939) as maximum V and is believed to be due to yielding of the grains parallel to the base (0001) during deformation. In other areas where the Wissahickon formation has been studied a similar arrangement is common and indicates probably the typical quartz orientation in a tectonite in reference to an



(a) (b)

FIGURE 4.—*Quartz diagrams of Wissahickon schist and Laurel gneiss*

(*a*) Wissahickon schist. Locality same as 2a. 400 average size quartz grains. Contours 1-2-3-4. Cleavage traces from Figures 2a and 2b. Mu = muscovite; Bi = Biotite.

(*b*) Laurel gneiss. Locality same as 3a. 200 large quartz grains. Contours 1-2-3-4-5-6-7-8. Main maxima in muscovite cleavage (Mu) and the intersection of biotite cleavages (Bi).

See Figure 2 for contour shading.

axis of rotation. The axis emerges about 10 degrees east of the center of the diagram and at the intersection of muscovite and biotite cleavage planes. The incompleteness of the quartz girdle is in harmony with the lack of a mica girdle.

Figure 4a was prepared in two stages of 200 grains each, and the maxima of the two partial diagrams reached 8 per cent. The total diagram is otherwise identical with each of the two partial ones. The present orientation shown in Figure 4a was thus fixed after 200 grains had been plotted.

In the Laurel gneiss (Fig. 4b) quartz axes are concentrated in a steep 8 per cent maximum which is in the muscovite cleavage plane. Since large and elongated quartz grains were measured, the author assumes that this orientation is due, in part at least, to the shape of the grains and not entirely to deformation and recrystallization. In order to avoid a concentration of axes of portions of one large, broken quartz grain in a false maximum, individual grains which were close together were counted as one. In addition, the dimensions of grains were measured, and the ratios plotted in Figure 5.

It is obvious that the concentration of points in Figure 5 is identical with the maxima in diagram Figure 4b. A close relation between grain shape and orientation thus becomes evident since those grains with the highest ratios (3:1 and above) are located in the principal maxima.

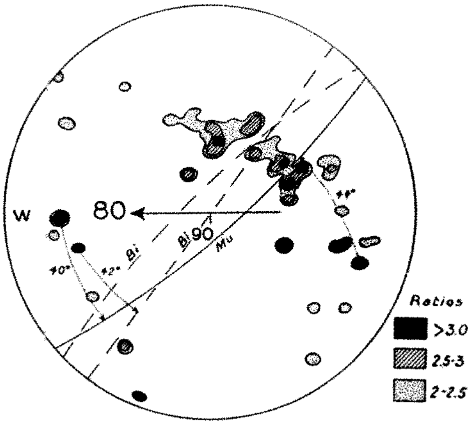


FIGURE 5.—*Quartz diagram of Laurel gneiss*

Locality same as 3a. Ratio of longest over shortest dimension of 100 quartz grains plotted in the axes points of Figure 4b. Solid and broken lines represent cleavage traces. Bi and Mu indicate traces of biotite and muscovite cleavage planes respectively.

Interpretation of quartz orientation.—All except two of the high ratio areas in Figure 5 coincide with important quartz maxima. The largest concentration is in the cleavage planes of muscovite (Mu) and biotite (Bi). These quartz grains are thus elongated parallel to the prism (1010) which in turn is very near or within the foliation plane. A second group of grains is concentrated in areas which are 40 to 45 degrees from the foliation planes (40, 42, and 44 degrees in Fig. 5).

Since all these quartz grains are broken into many elongated fragments, the fracturing seems to have taken place largely along "cleavages" which can be identified as mineralogical directions: base, prism, and rhombohedron. Their later orientation in the Laurel gneiss took place predominantly according to shape. The large quartz grains indicate flow structures, and their orientation appears to represent flow lines. Similar orientations are well known from many igneous masses in which fragments or elongated crystals are orientated parallel to flow lines or planes (Balk, 1937). The author believes that the fragmentation of quartz took place during an early stage of granulation and that the fragments were oriented later largely according to form. Fragmentation took place along quartz cleavages and parallel to common crystal directions.

TEXTURAL RELATIONS

The texture of the Laurel gneiss is quite different from that of the schist from the Wissahickon formation. The schist is a coarse, foliated rock which consists of bands of quartz and oligoclase alternating with layers of coarse, well-oriented micas. The gneiss, on the other hand, is a fine-grained, sugary rock with a crude alignment of biotite and muscovite. The micas are scattered throughout the rock and are not localized in bands as in the schist. The texture of the schist, therefore, is that of a strongly metamorphosed rock, whereas the megascopic texture of the gneiss appears more like that of an igneous type.

GRADATIONAL CONTACTS

Wherever exposed, the boundary between the Laurel gneiss and the Wissahickon formation is gradational; in many places it is difficult to establish. In the vicinity of locality 1 (Pl. 1), for example, the Wissahickon schist grades into typical Laurel gneiss within a distance of about 300 feet. A similar relation is shown in a traverse across the contact from locality 28 to locality 33 along the Patuxent River (Pl. 1). Here the rock changes gradually from a fine, granular, poorly foliated type (typical Laurel gneiss) into the coarse, foliated mica schist and interbedded quartzite of the Wissahickon formation. The transitional relation may also be studied at locality 56 (Pl. 1).

REMNANTS

Another important relationship is shown by the presence of remnants of schist and quartzite in the Laurel gneiss. The schist remnants show irregular boundaries across which there is a gradual transition from schist into gneiss. Even under the microscope it is difficult to determine where one ends and the other begins. The change is entirely textural since the mode of the schist remnants and that of the gneiss are essentially the same.

ORIGIN OF THE LAUREL GNEISS

GENERAL STATEMENT

Only two explanations of the origin of the Laurel gneiss appear plausible. It may be the result of "granitization" of the Wissahickon formation by an intrusive magma or it may have originated by metamorphism of the Wissahickon formation.

HYPOTHESIS OF GRANITIZATION

Cloos interpreted the Laurel gneiss as a migmatite—the result of granitization of the Wissahickon formation. He believed that the latter was actually penetrated, dissolved, and assimilated by intrusions of granitic magma. Cloos' hypothesis was based upon field observations, and the Laurel gneiss, as mentioned above, has many of the megascopic features of an igneous rock. It might also be assumed that the Wissahickon formation was granitized through the introduction of material by solutions from a granitic magma rather than by the magma itself.

Some arguments in favor of granitization of the Wissahickon formation are as follows: (1) The normal Laurel gneiss has the appearance of a granular igneous rock with flow structure; (2) it has a chemical composition somewhat similar to that of a true granite (Table 2); (3) it contains abundant fragments of schist and quartzite from the Wissahickon formation, in some places so many that the gneiss looks like a hybrid rock; and (4) a granitic intrusive (Guilford granite) bounds the Laurel gneiss on the north.

A number of facts, however, individually suggest that the Laurel gneiss is not the result of granitization. None of these arguments is, in itself, a definite proof, but when considered together they indicate that the gneiss was not formed by granitization.

(1) Sharp intrusive boundaries between the Laurel gneiss and the Wissahickon formation have never been found. This means that the former could not have been formed by the injection of granitic magma into the Wissahickon formation, otherwise one would expect to find at least a few dikes and sills penetrating the schist. On the other hand, this argument does not preclude the possibility of granitization by solutions since this process would normally produce gradational boundaries.

(2) The Laurel gneiss is, as a rule, finer-grained than the Wissahickon formation. If the Wissahickon formation had been granitized, one would expect the resulting rock to be at least as coarse as, if not considerably coarser than, the original rock. Recrystallization is an important feature of granitization, and this increases grain size.

(3) No mineral occurs in the gneiss that has not been found in the Wissahickon formation in the surrounding area. Furthermore, those minerals in the former are exactly the same optically (and, therefore, chemically) as those in the schist and quartzite. It would be a strange coincidence indeed if the minerals of an igneous intrusive, or those introduced by solutions, were exactly the same as those of the country rock. It is admitted that this might be possible if considered in the light of the principle of mineral facies. Thus an igneous rock and a meta-

morphic rock may contain the same mineral assemblage providing that both had the same original chemical composition and that both formed under similar temperature-pressure conditions. However, it seems very unlikely that the intrusion would have crystallized under exactly the same temperature-pressure conditions as those under which the Wissahickon formation was metamorphosed. It appears, therefore, that the similarity in mineral composition between the Laurel gneiss and the Wissahickon formation is significant evidence against the hypothesis of granitization.

(4) The so-called Guilford "granite" (really a granodiorite), which is the only exposed rock close enough to have caused the granitization, could not have had anything to do with the origin of the Laurel gneiss for several reasons. In the first place, the Guilford granite is younger than the gneiss. Numerous dikes of it cut the gneiss near locality 1 (Pl. 1), and a wide dike occurs in the stream at locality 60 (Pl. 1). Their boundaries are always sharp and distinct and are never gradational or characterized by stringers and inclusions. Coarse, pink pegmatites from the Guilford granite cut the gneiss along clean, sharp boundaries in several localities. Secondly, the Guilford granite contains a high percentage of potash feldspar. A mode of this rock (8) is shown in Table 1. In all the sections of gneiss examined, however, no potash feldspar has been found. In the third place, the Guilford granite is texturally unlike the gneiss. It is a light-gray, nonfoliated, biotite rock with a grain size of from 1 to 2 millimeters. The gneiss, it will be remembered, is faintly foliated and finer-grained.

Because of these facts, and because the Laurel gneiss can be explained more satisfactorily in another way, the writer does not believe that the hypothesis of granitization of the Wissahickon formation either by an intrusive rock or by solutions can be accepted.

HYPOTHESIS OF METAMORPHISM

Introduction.—A completely satisfactory explanation of the origin of the Laurel gneiss must take into account the following pertinent facts: (1) The mineralogical and chemical compositions of the Laurel gneiss are essentially the same as those of the Wissahickon formation; (2) wherever the contact can be studied there is a complete textural gradation from the one rock type into the other; (3) the gneiss is younger than the Wissahickon formation and contains abundant fragments (remnants) of schist and quartzite; (4) the gneiss has undergone both granulation and recrystallization, and the structure and fabric indicate that it has flowed as a plastic mass. After considering these facts, the writer is convinced that the Laurel gneiss originated by the granulation, flowage, and recrystallization of the Wissahickon formation. Granulation took place under shearing stress in the early stage of the process when the rocks were cold and dry; flowage and recrystallization occurred later under conditions of higher temperature and abundant water. It is believed that the complete transformation took place without appreciable addition or loss of material.

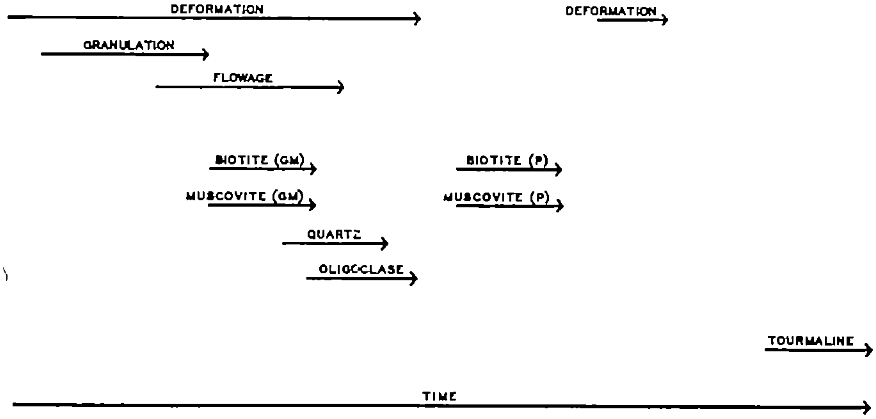


FIGURE 6—*Graph showing time relationships between the various stages in formation of the Laurel gneiss*

GM = groundmass, P = porphyroblasts

The various stages in this process will be described below in detail. The time relationships between the different stages is shown graphically in Figure 6.

Granulation.—The evidence for granulation is now largely obliterated by recrystallization. That this was once an important process in the formation of the Laurel gneiss, however, is shown in several ways:

(1) The typical Laurel gneiss is finer-grained than the schist of the Wissahickon formation. Since recrystallization ordinarily tends to increase grain size, it would suggest that the schist was crushed and granulated before it was recrystallized.

(2) The strongest and most direct evidence of granulation is furnished by the universal straining and breaking of quartz grains in the gneiss. (*See* petrographic description and discussion of fabric.) Many large crystals of quartz have been strained almost to the breaking point, whereas others have actually broken into smaller fragments along irregular cracks. Many of these fragments have been displaced relative to the large crystal, but some are still sufficiently close in optical parallelism so that there can be no doubt that they represent parts of a once complete crystal.

In the interior of the gneiss body the granulation of the quartz is more or less complete, but in the transition zone between the Laurel gneiss and the Wissahickon formation there are numerous large grains of quartz which have not been fully broken down (Pl. 6, fig. 4).

(3) At least one quartzite remnant has been faulted (Pl. 2, fig. 2), and the fault plane healed by later flow and recrystallization.

The writer believes that the granulation took place during the early stages of transformation by the complete mechanical destruction ("Durchbewegung") of the Wissahickon formation. In these early stages the rocks were cold and dry and yielded by granulation and crushing. The large quartz grains in the schist were strained and broken into smaller grains. The oligoclase crystals were probably also crushed and granulated in a similar manner although little evidence of this can be seen in the gneiss today. The micas were bent and twisted, and both they and the garnets were broken into smaller fragments. Thus, by this process the coarse Wissahickon formation was transformed into a finer cataclastic rock. Locally, where stress conditions were less severe, the schist and quartzite were not completely granulated, and remnants of these are now preserved in the gneiss.

Flowage.—With the increase of temperature and solutions, the rock mass began to yield by flowage as well as by granulation. Individual grains of quartz, oligoclase, and mica moved about relative to one another, partly by slipping along crystal boundaries and partly by plastic deformation of grains. Mica crystals became crudely oriented in response to differential movements in the mass. Many remnants of schist and quartzite which were not previously granulated were twisted about into random orientations. Continued granulation accompanied this process, but, as the temperature rose and water increased in amount, flowage became the dominant mode of adjustment to the stress, and crushing and granulation became less important.

Rock flowage did not cease at the end of the period of crushing and granulation but continued on well into the period of recrystallization. This is certain because many recrystallization features show evidences of this flowage: (1) Mica segregations are pulled out into streaks and bands; (2) the foliation, which is caused by roughly oriented recrystallized micas, is locally thrown into complex folds; (3) the foliation wraps around many quartzite remnants in a manner suggesting that it actually flowed about them; and (4) the fabric of the gneiss, which is at least partly a result of recrystallization, shows unquestionable evidence of flowage. The late stages of rock flowage and the early stages of recrystallization were thus contemporaneous.

Recrystallization.—After granulation had largely ceased, but while some flowage was still in progress, recrystallization of the crushed rock began. Although stress still operated, it was becoming relatively less important, and heat and water were becoming the dominant factors. Large grains grew at the expense of small ones, strained crystals tended to dissolve and unstrained ones to take their place, and micas grew in directions of least resistance or were oriented by flowage so that the rock took on a faint foliation. Thus the rock lost much of its crushed and granulated appearance.

Recrystallization of the essential minerals may be divided into two stages: (1) formation of minerals of the groundmass, and (2) formation of porphyroblasts.

In the first stage quartz, plagioclase, and the oriented micas formed. It is clear that the quartz and plagioclase are essentially contemporaneous in their crystallization since both minerals are mutually interlocked by sutured boundaries. The oriented micas formed either slightly before or during the recrystallization of the quartz and plagioclase. Their orientation is due probably to two factors—*i.e.*, crystallization under stress and rock flowage.

The porphyroblasts formed during a later stage of recrystallization, after the stress had ceased entirely. The muscovite and biotite porphyroblasts cut across the foliation of the gneiss and inclose poikiloblastically crystals of apatite, magnetite, zircon, quartz, plagioclase, and small crystals of mica. Figure 2 of Plate 6 shows a muscovite porphyroblast with inclusions of poorly oriented biotite. It has already been pointed out that the muscovite and biotite porphyroblasts are mutually intergrown in many cases, suggesting simultaneous crystallization. They show undulatory extinction due to straining and prove that a period of renewed stress followed their formation. These events were followed by the crystallization of scattered porphyroblasts of tourmaline, some of which appear to cut directly across mica porphyroblasts. The position of the garnets in the sequence is not entirely clear. Most of them appear to be remnants of coarse porphyroblasts which formed originally in the schist and were not completely broken down by granulation. A few of them have undergone some recrystallization.

It has been shown that the musovite and biotite porphyroblasts have the same optical properties as the fine, oriented muscovite and biotite in the groundmass, and that their chemical compositions are, therefore, probably the same. It has also been pointed out that the porphyroblasts originated later than the groundmass micas through the medium of hot aqueous solutions. There is no evidence whatever that any material was added to or removed from the gneiss by these solutions; they appear to have acted simply as agents for recrystallization by dissolving the groundmass micas and depositing them as porphyroblasts. Barth (1936, p. 805-806) has advocated a somewhat similar explanation for the muscovite metacrysts in the porphyroblastic muscovite schist of Dutchess County, New York.

Although granulation, flowage, and the early stages of recrystallization were confined to the area of the Laurel gneiss, the formation of late porphyroblasts extended out into the Wissahickon formation. This suggests that in the later stages of recrystallization solutions had become effective over a large area.

The segregations of mica porphyroblasts with their surrounding areas of cream-colored rock also belong to this late stage of recrystallization. It is thought that they originated in essentially the same way as the scattered porphyroblasts. In this case, however, the process was locally more complete. Practically all groundmass micas were dissolved from small areas of the gneiss, thus leaving the rock cream-colored. The dissolved material was then reprecipitated as porphyroblasts around specific centers.

Agencies of metamorphism.—The agencies of metamorphism were: (1) shearing stress, (2) heat, and (3) solutions (water).

The stress causing granulation and flowage is believed to represent a late phase of the regional stress which produced the foliation in the Wissahickon formation. The general conformity of the foliation in the Laurel gneiss with that in the Wissahickon formation suggests that the late stress phase acted in essentially the same direction as the main stress. We have no way of knowing the exact shape of the gneiss body since much of it is now covered by Cretaceous sediments. However, in general, it is long and narrow and trends northeast-southwest parallel to the regional strike. This suggests that the gneiss developed along a broad fault zone or shear zone where crushing and granulation were most intense.

The heat necessary for flowage and recrystallization is believed to have originated in three ways: (1) heat from the geothermal gradient, (2) heat generated mechanically by crushing and granulation, and (3) heat brought up by conduction and by solutions from a magma below.

The temperature at which the gneiss formed can be determined in a general way if two assumptions are made: (1) The Wissahickon formation was overlain by all the Paleozoic formations at the time of its metamorphism (Permian?), and (2) the gneiss formed shortly after or during the latest phase of the metamorphism of the Wissahickon formation before the load was appreciably reduced. Ernst Cloos (personal communication) estimates that the Paleozoic column above the gneiss must have had a minimum thickness of 30,000 feet or 9 kilometers. According to Adams (1924, p. 468, Fig. 2), the temperature at this depth, resulting from the geothermal gradient, is approximately 250° C. If we add to this the heat generated by crushing and granulation, and that introduced by solutions and by conduction from a magma below, we might at first conclude that the temperature was several hundred degrees. However, there is nothing to indicate that the temperature was exceptionally high. For example, no portion of the rock shows any evidence of fusion. Furthermore, recrystallization was not complete since much of the quartz is still strained. More specifically, the absence of orthoclase suggests that the temperature during recrystallization was not above 340° C. (Grout, 1932, p. 441). Roughly, the temperature during metamorphism probably ranged between 250° and 340°.

The writer believes that water played an important role in the flowage and recrystallization of the Laurel gneiss. This is suggested in two ways: (1) Contorted foliation indicates that the gneiss was, at least locally, a very plastic mass The importance of water in increasing the plasticity of rocks is well recognized. (2) The abundance of porphyroblasts and the occurrence of irregular segregations of mica suggest that water penetrated the gneiss thoroughly during the final stage of transformation. (*See* section on recrystallization.) The water is thought to have had two sources: (1) water liberated from the micas during the metamorphism, and (2) water rising from below along a fault zone or shear zone. The presence of tourmaline as late porphyroblasts in both the gneiss and the schist suggests an igneous body at depth from which both the water and heat emanated. Except for scattered crystals of tourmaline, there is no evidence whatever that material was added or subtracted during recrystallization. Both the Laurel gneiss and the Wissahickon formation have essentially the same chemical compositions. This suggests that the water promoting recrystallization was relatively pure.

PETROGENETIC IMPORTANCE OF THE LAUREL GNEISS

The problem of the Laurel gneiss raises a question of fundamental importance in the science of petrogenesis. Here is a rock whose general characteristics and relation to the surrounding rocks indicate that it was formed simply by the metamorphism of the Wissahickon formation without the addition or subtraction of appreciable material. And yet this rock possesses most of the features of a typical igneous rock. It has the structure and megascopic texture of an igneous rock, and a chemical composition similar to that of granite. It likewise contains segregations which are not unlike those found in many igneous bodies. Ernst Cloos has shown that the fabric of the gneiss is somewhat similar to that of an igneous rock which has flowed. Furthermore, the gneiss contains fragments of schist and quartzite which might easily be interpreted as inclusions engulfed in a magma. In fact, where these remnants are abundant, the Laurel gneiss closely resembles a migmatite, and the writer, therefore, suggests that it be called a "pseudomigmatite."

There are two differences, however, between the Laurel gneiss and an igneous rock: (1) The mineral composition of the gneiss is not that of any typical igneous rock (but is the same as the wall rock), and (2) the gneiss shows gradational contacts with the adjacent Wissahickon formation.

The chief mineralogical difference between the Laurel gneiss and a typical granite is that muscovite occurs instead of orthoclase. However, if the temperature of metamorphism had been higher (probably above 340° C.), orthoclase might have formed instead of muscovite. Then the gneiss would have had not only the appearance of a granite and a somewhat similar chemical composition, but also the mineral composition of a true granite.

Under somewhat different conditions of metamorphism the Laurel gneiss might also have developed intrusive contacts. Any agent which could increase the plasticity or mobility of the rock might cause it to flow more readily and thereby intrude the surrounding rocks. Intense stress would, without doubt, be an important agent. Then too, Griggs (1940) has clearly demonstrated that water is of fundamental importance in rock flowage. The effect of high temperature is somewhat debatable, but both field data (Grout, 1932, p. 390) and some experimental data (Adams, 1912, p. 111-112) indicate that, in general, rocks yield more readily at high temperatures. Let us suppose that during transformation the stress had been more intense, the temperature had been higher, and solutions had been more abundant. Under these conditions the gneiss would have been more plastic and more mobile and might easily have penetrated cracks and crevices in the surrounding country rock to form dikes and sills with true igneous boundaries.

It appears, then, that the Laurel gneiss might have had all the features of a true igneous rock if the agencies of metamorphism had been more intense. Under these conditions it would have been impossible to distinguish it from an igneous rock by any criterion.

In recent years much has been written on the subject of igneous-looking rocks which have presumably been formed by some process other than the crystallization of magma (Goldschmidt, 1921; Sederholm, 1926; Eskola, 1932, 1933; Gilluly, 1933; Barth, 1936; Wegmann, 1936; Anderson, 1937; C. A. Chapman, 1939, p. 167-170; Postel, 1940). For example, "syntexis" is the process by which crystalline or sedimentary rocks are either penetrated by magma or altered by magmatic solutions until, in some cases, they look like true igneous rocks. "Ultrametamorphism" is the supposed process by which a solid rock becomes so intensely metamorphosed that it is fused, either selectively or completely, and then crystallized again to produce an igneous-looking rock. Many variations of these processes have been described in the literature. The difficulties in distinguishing between igneous-looking rocks produced by such processes and true igneous rocks which have crystallized from magmas are now well recognized in petrogenesis. They have been briefly summarized in a recent paper by Grout (1941).

Like some of the syntetic rocks, the Laurel gneiss has most of the features of an igneous rock and in the field can easily be mistaken for one. The gneiss is apparently not a syntetic rock, however, since there is no evidence whatever that any appreciable amount of magmatic material was introduced during its formation. Nor is the gneiss a product of ultrametamorphism because the evidence cited previously in this paper indicates that no part of it has ever been remelted. Instead, the facts strongly suggest that the gneiss is a true metamorphic rock which has developed an igneous appearance simply by granulation, flowage, and recrystallization.

The writer is not aware that any large body of igneous-looking rock has ever been explained in such a way as this, at least not in recent literature. To be sure many igneous-looking rocks are interpreted as products of metamorphism, but in these cases the term "metamorphism" usually involves either the addition of igneous material to some older rock (syntexis), or the fusion and recrystallization of an older formation (ultrametamorphism). A fundamental, petrogenetic problem is thus suggested by the Laurel gneiss since its origin appears to be unique. It would be interesting to know whether there are other igneous-looking rocks that might have originated in the same manner.

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EXPLANATION OF PLATE 6

PHOTOMICROGRAPHS OF TYPICAL WISSAHICKON SCHIST AND LAUREL GNEISS

Figure 1.—Typical Wissahickon schist from locality 1 (Pl. 1). Foliation caused by layers of well-oriented mica alternating with layers of equidimensional quartz and plagioclase. Note straining in quartz. Diameter of field 5 millimeters, crossed nicols.

Figure 2.—Typical Laurel gneiss near locality 28 (Pl. 1). Note inequigranular texture and irregular strained quartz grains with sutured boundaries. In contrast with typical Wissahickon schist this rock has a poor foliation determined by small, crude aligned mica grains and somewhat elongate quartz crystals. Note large muscovite porphyroblast with inclusions of poorly oriented biotite. Diameter of field 5 millimeters, crossed nicols.

Figure 3.—Laurel gneiss, 500 feet west of locality 20 (Pl. 1), with large, highly strained quartz lump Note that lump is intercrystallized with surrounding gneiss, proving it is a remnant and not introduced later Note also the granulated, recrystallized appearance of the rock. Diameter of field 5 millimeters, crossed nicols.

Figure 4.—Photomicrograph of large quartz crystal strained and broken into fragments. The rock is from the transition zone between the Laurel gneiss and the Wissahickon formation at locality 28 (Pl. 1). Diameter of field 5 millimeters, crossed nicols.

Figure 5.—Large quartz grain shown in figure 4 of this plate. The original grain, surrounded by the heavy black line, is now broken into many fragments Arrows and figures indicate attitude of c axes within individual fragments.

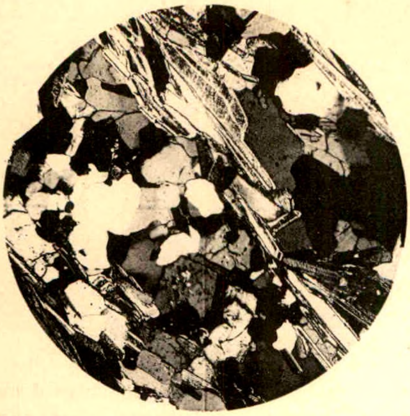

FIGURE 1

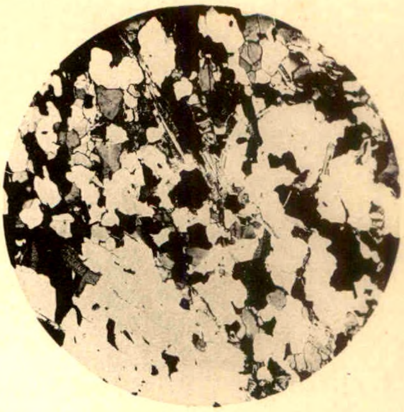

FIGURE 2

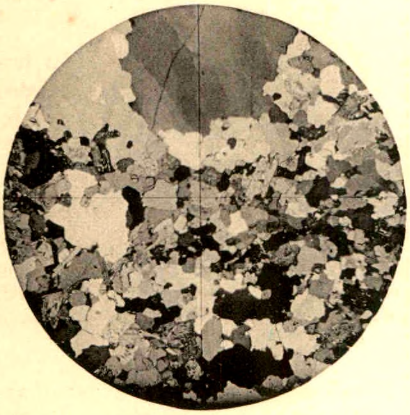

FIGURE 3

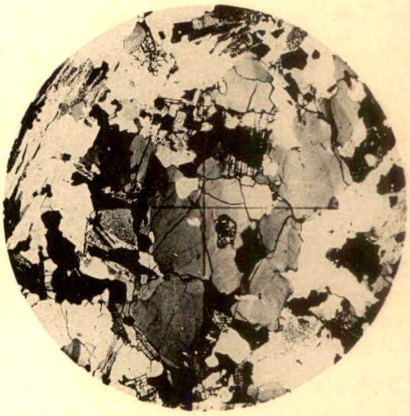

FIGURE 4

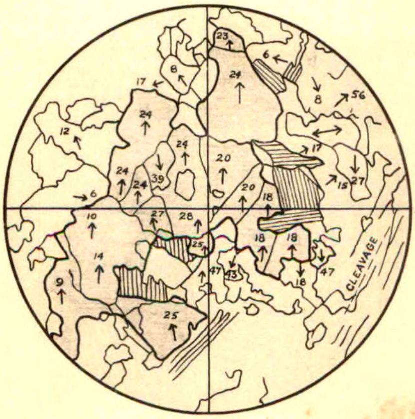

FIGURE 5

PHOTOMICROGRAPHS OF TYPICAL WISSAHICKON SCHIST AND LAUREL GNEISS

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NATIONAL RESEARCH COUNCIL AND CO-OPERATION IN GEOLOGICAL RESEARCH

BY WALTER H. BUCHER

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INTRODUCTION

All research, in the end, is the product of individual initiative and effort, so much so that we are apt to overlook the growing need for co-operation among scientific workers in the ordering and making available of results and in the planning, financing, and execution of investigations which in scope exceed the powers of the individual. The time is past when a Dana or a Zittel could collate all knowledge of one field in a handbook, nor are we inclined individually to spend "too much" of our time "compiling". And the time has come when new facts and new roads to understanding lie either in less accessible parts of the globe or in the "borderlands" of science which are no less difficult of access to the individual. Co-operation has accordingly become widespread among men in one field of science as well as across the borderlands between different sciences.

This trend toward co-operation was stimulated greatly when in 1919[1] the National Research Council was transformed from an instrument of war to one of peace, charged with the task of promoting "research in the mathematical, physical, and biological sciences, and in the application of these sciences to engineering, agriculture, medicine, and other useful arts. . . ." (National Research Council, 1933.) The charge was a new one as applied to the whole field of a science. How new it was is shown by the efforts that were made in each Division to define the scope of the Council's work, as illustrated in geology by E. B. Mathews' *Outline of the scope of the National Research Council Division of Geology and Geography* in the chairman's report for 1921, N. M. Fenneman's paper on the functions of the Division (1922), and the list of some 48 possible projects discussed in the first five divisional meetings (Annual Report, 1922, esp. Appendix B). Much has since been accomplished, more than is generally recognized. A brief survey of the lines of action that have proved effective would be worth while in itself now, after two decades of results. The immediate reason, however, for giving such a summary at this time lies in the growing interest of The Geological Society of America in a research "program."

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[1] The executive order is dated 1918, but the reorganization took place in 1919

The discussion that is now going on in this Society is much like that which led to the consolidation of opinion concerning a research program in the American Association of Petroleum Geologists some 14 years ago. At the meeting of the Association in San Francisco in 1928, Heald began a paper on *Research and the American Association of Petroleum Geologists* with questions such as are now being debated among members of the Geological Society, as for instance, "Should it [the Association] initiate work or should it merely support work that has been started by some other organization?" As with us now there existed then "some uncertainty as to the scope and functions of the National Research Council and the American Petroleum Institute, the two organizations that are most actively fostering research in problems of petroleum geology" (Heald, 1928, p. 939).

The first question had been answered already with firmness and vision by W. E. Wrather in the progress report of the Committee on Research, presented at the tenth meeting of the Association in 1925. He wrote:

> "In view of conditions prevailing in the oil industry, the functions of a Research Committee should be (1) to visualize and delineate particular problems of needed research; (2) to broadcast and popularize these problems to the men who are constantly in the field, and encourage them to gather and transmit information of the type needed for the prosecution of such problems; (3) to locate men who are particularly interested in these problems, and (4) to supply these men the necessary data, as well as proper facilities and financing, to enable them to make the most of the problem in hand." (Wrather, 1925, p. 2.)

This policy has prevailed and has been amply justified by results.

The uncertainty concerning the relations with the National Research Council was justified at the time. The latter's "Committee on Studies in Petroleum Geology" was organized in the same year (1923) as the Committee on Research of the American Association of Petroleum Geologists, and all its members were also members of the Association. They were, as Heald said, "really complementary groups. They have co-operated closely ever since their inception, and there is no clear-cut line of demarcation between their functions." Several projects were undertaken jointly. When the geological map of Oklahoma was being prepared by the State Survey of Oklahoma in co-operation with the U. S. Geological Survey, the oil geologists and oil companies of the State contributed through the National Research Council (U. S. Geol. Survey, 46th Ann. Report, 1925, p. 29). When, about 1926, John D. Rockefeller placed at the disposal of the American Petroleum Institute the sum of $250,000 for research on problems in petroleum geology and the Universal Oil Products Company of Chicago donated an equal amount, the Institute turned to the National Research Council for advice and co-operation in the selection and execution of suitable projects. At joint meetings of the committees of the National Research Council and the Association, problems were discussed and projects formulated and cast into suitable form for approval by the Institute. This set a fine example of co-operation in scientific research between two national organizations. Since then the Association has become quite independent and has, in fact, set a standard in the planning and development of co-operative research projects. A close liaison in personnel and a regular exchange of information is of course being maintained, which is fruitful to both the Association and the National Research Council.

In The Geological Society of America, the idea of a "research program" is only now taking hold. It may be defined as the desire to invite the planning of major projects which the membership recognizes as yielding results of basic value to our science, to which it can point with pride and which it is therefore willing to support liberally.

The idea took form officially in 1936 when Fenneman challenged the Geological Society to think of research not only in terms of projects offered at random by individual members, but also as something that in some of its larger aspects requires deliberate collective planning for truly significant and far-reaching results. In response to his words a "Committee on Research Program" was created but was given only advisory function. In 1939 the committee still reported that "except for the work of advisory committees, a possible larger scope of initiative on the part of the Society has not been called for." In 1940 the Budget Committee suggested differentiation between normal miscellaneous projects and larger projects designed to advance geological knowledge along basic lines and tentatively earmarked the round sum of $10,000 as available for the latter. The response, both oral and in letters addressed to the chairman of the Committee on Research Program, showed that to many members the thought was still new, that they might look collectively at the needs of the science and plan deliberately to fill them by inviting and supporting investigations of large import. This habit of "looking collectively at the needs of the science" is essential for the healthy growth of a research program. That, however, has been and is the task of the Division of Geology and Geography. Its experience should be of value. It is best conveyed through a survey of the more important results of the division's labors. This is given in the following pages at the request of the Council of the Geological Society, made upon recommendation from the Society's Committee on Research Program, together with suggestions of promising lines of research that may help crystallize the concept of a research program within The Geological Society of America.

From a practical point of view, the National Research Council may be characterized concisely as (1) a forum where men of diverse affiliations are brought together to discuss the needs of the science and to arouse interest and co-operation in the development of new fields of knowledge or neglected old ones; (2) an official body that may effectively ask the co-operation of many individuals in research, and especially also in the creation of tools for research which other organizations are not interested in or not capable of creating; and (3) an organization that stands above suspicion of profit-making, in whose hands government agencies, corporations, and private individuals may place funds for use in research.

By bringing together in one organization men from all branches of natural science, unique facilities are created for promoting interdivisional co-operation which are difficult to achieve in other ways. The results of this type of work will be taken up first, not only because it includes spectacular achievements, but also because it serves to widen the angle of vision in a manner useful in retrospect and outlook along purely geological lines.

BORDERLINE FIELDS BETWEEN GEOLOGY AND OTHER SCIENCES

GEOLOGY AND PHYSICS

The largest single contribution to both physics and geology was made by the Division of Physical Sciences with the co-operation of the Division of Geology and Geography and the American Geophysical Union. In 1926, Dr. J. S. Ames as chairman of the Division of Physical Sciences set out to fill the need of systematic treatises on geophysics in the English language. The resulting series of volumes is a tribute to him and his successors and above all to the members of the large committee to which the task was entrusted. All nine volumes of the series, *Physics of the Earth,* have appeared. The first six were published as Bulletins of the National Research Council, the others by McGraw-Hill Book Co. as follows:

I. Volcanology (77, 1931); II. The Figure of the Earth (78, 1931); III. Meteorology (79, 1931); IV. The Age of the Earth (80, 1931); V. Oceanography (85, 1932); VI. Seismology (90, 1933); VII. Internal Constitution of the Earth (1939); VIII. Terrestrial Magnetism and Electricity (1939); IX. Hydrology (1942).

In the spring of 1936 Dr. E. S. Bastin, as chairman of the Division of Geology and Geography, sent letters of inquiry to about 300 geologists and geographers asking them which lines of research were regarded as especially promising in the fields of geology and geography on the one hand and in the borderlands between geology and physics and chemistry on the other. A digest of the replies to the first question was published in 1936 under the title: *Suggestions concerning desirable lines of research in the fields of geology and geography.*[2] With it a comparable list of subjects should be consulted which was secured from Fellows of the Geological Society through a circular letter sent out by the Society's Committee on Research and published in the Proceedings for 1937 (Vaughan, 1938).

For the sifting and appraising of the replies concerning borderland problems a special, interdivisional committee was created under the chairmanship of T. S. Lovering. Its report (1938) is rich in trenchant formulations of problems and important references. It led directly to the organization of two interdivisional committees with specific functions. One was to fill the wide-felt need for a concise and up-to-date collection of the physical constants of geologic materials. Under the chairmanship of Francis Birch, this important work has been completed and published by the Geological Society (Special Paper 36, 1942). The other Committee, on Density Currents, with Herbert N. Eaton as chairman, grew out of the interest in bottom currents of "standing" water bodies which had been aroused by Daly's brilliant exposition of his hypothesis concerning the origin of submarine canyons. A large amount of quantitative data was collected in Lake Mead concerning all essential factors in the hydrology of the lake. A bulky report, containing the data in tables and graphs, has been completed and is available in typewritten form at a number of places in the United States, including the Division's office. A digest of the data is being prepared for publication.

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[2] For the latest list of *Publications of the National Research Council, Washington, D C, of interest to geologists and geographers,* write the Secretary of the Division of Geology and Geography, National Research Council, 2101 Constitution Avenue, Washington, D C.

In 1923, six years after Barrell had refuted Becker's ill-considered denunciation of the uranium-lead method of measuring geological time and had awakened American geologists to the possibilities of this method (Barrell, 1917), a Committee on Measurement of Geologic Time by Atomic Disintegration was organized at Lawson's suggestion with A. C. Lane as chairman. Beginning with a few pages in 1924, the committee's report quickly grew to mimeographed volumes giving annual summaries of work accomplished (annotated bibliographies) and pages of critical comments and stimulating outlook toward needed investigations. Even though it was not organized as an interdivisional committee, it comprised from its beginning the leading workers in the field of radioactivity in America and maintained regular correspondence with those in foreign countries. It has made possible much costly work through funds secured from a number of sources, including rather large sums from anonymous donors. In the later years the Geological Society has supported experimental work to the extent of more than $25,000. We have a right to be proud of what has been accomplished under our auspices. Few realize as yet in how many directions the newest development in this field, Goodman's and Hurley's work in the helium-content of heavy minerals, promises to shed new light. The utilization of the radioactive record for geological purposes actually has barely begun. The Advisory Committee on Research Program, accordingly, recommended last October to the Council of the Geological Society "that investigations of the radioactive measure of geological time be continued as a part of the Society's program which is given major emphasis."

The support which the Society has given to this work justifies reference to it as part of the Society's program. It illustrates one of several ways in which the units of a research program come into being.

It is to be expected that, in general, units of this program which require for their formulation the co-operation of organization physicists and geologists should take form and find impetus in the American Geophysical Union. This organization was established in 1919 as the American Committee of the International Union of Geodesy and Geophysics under the executive control of the Committee on Geophysics of the National Research Council.

> "The objects of the Union are to promote the study of problems concerned with the figure and physics of the Earth, to initiate and coordinate researches which depend upon international and national cooperation, and to provide for their scientific discussion and publication"

The growth of the "Transactions" through the years offers a remarkable example of the good that comes from bringing together men from two fields of science who not only speak the same language but whose intellectual curiosity extends far enough to embrace the other's range of vision.

The records of the Division of Geology and Geography contain more than one example to show that without team work across the borderline the geologist can achieve little along geophysical lines of inquiry. In 1923, for example, a "Committee on Detection of Subsurface Structures by Sounding" was discontinued because the circumstances under which investigations in this subject were conducted at the Bureau of Standards made it impossible for the committee to procure information at that time. The geologists apparently were not then thinking of other sources of information and co-operation.

A "Committee on Sonic Sounding and Oceanographic Thermographs" was organized in 1922 at the suggestion and under the chairmanship of T. Wayland Vaughan. It was renamed in 1925 "Committee on Submarine Configuration and Oceanic Circulation."

> "The principal duty of this committee was to obtain information for distribution to geologists and geographers, and also to put the members of the Division in a position to lend support to investigations of the two kinds mentioned." (Annual Report, 1923, Appendix "H".)

At a time when few American geologists knew anything about investigations in physical oceanography, and years before the International Council for the Exploration of the Sea began to publish such information for oceanographers, this committee did most valuable service through its annual reports, the last two of which comprised over 100 pages each. But it is significant that among the duties of the committee no mention is made of deliberate planning of work designed to secure knowledge of basic value to geology and enlisting the co-operation of competent individuals and organizations in its execution. Such co-operation in planning and doing requires that the right men be brought together. Chiefly to make this possible, the functions of this committee were transferred in 1932 to the Geophysical Union. In this new environment such investigations as Ewing's seismic studies on the structure of the continental shelf and Piggot's and Stetson's studies on Atlantic sediments were vitalized if not conceived. (*See, e.g.*, Am. Geophys. Union, 1932.) The same may be said, *e.g.*, of Woollard's magnetic and gravitational surveys; Birch's and Slichter's studies on the velocity and propagation of seismic waves; and Grigg's work in mechanics of rock deformation under high pressures and temperatures.

The energy that is needed for such studies is drawn from two soils: that of geophysics which provides the methods of geodesy, oceanography, seismology, and physics of materials; and that of geology which provides the comprehensive body of knowledge from which the demand for new information springs. The work is done by men who are members of our Society. Their investigations are ours and fit beautifully into the requirements of a "research program" as stated above:

> "major projects which the membership recognizes as yielding results of basic value to our Science, to which it can point with pride, and which it is therefore willing to support liberally."

GEOLOGY AND CHEMISTRY

Geochemical problems have been less in the limelight in the last two decades than those of geophysics. Many have been under investigation for decades in the Geophysical Laboratory. Some that are no less pressing are waiting to be studied. Fleischer and Morey's work on the *Equilibrium between vapor and liquid phases in the system* CO_2-H_2O-K_2O-SiO_2 is a good example of an important investigation that was initiated by a small group within The Geological Society of America. The problem and its method of attack were formulated in 1935 at conferences between A. L. Day, G. W. Morey, C. H. Behre, Jr., and T. S. Lovering. The work was carried on in the Geophysical Laboratory. It was financed for two years by grants from the Penrose Bequest and for a third year by the Carnegie Institution of Washington. The results were published in 1940.

The report of Lovering's Interdivisional Committee previously referred to contains valuable suggestions. Since its publication, steps have been taken to secure action along some lines.

At a two-day meeting in October, 1940, called for the purpose, the Committee on Problems of Ore Deposits, with Lovering as chairman, analyzed in detail the whole range of problems connected with the physical chemistry of ore deposits. The "Summary of problems suggested" as a result of the joint counseling of five chemists, one physicist, and ten geologists is an important document (Annual Report 1940-1942, Appendix J). The committee went further, however. Having formulated the problems, it proceeded to decide on what to begin in a systematic series of investigations designed to supply vital basic data. The water-sulphide systems proved both feasible and basic. A first unit in this planned series was chosen: the study of the solubility of metallic sulphides in water at elevated temperatures and pressures. As before, the Geophysical Laboratory has consented to provide space, light, and power for the research worker, and, above all, its unique facilities and the invaluable advice of its staff. The Geological Society has agreed to finance this unit of the investigation.

GEOLOGY AND BIOLOGY

In contrast with this working hand-in-hand of geologists with physicists and chemists, the lack of systematic joint counseling of paleontologists with biologists is conspicuous. There have been no interdivisional committees, either of Geology or of Geography and Biology, in the National Research Council. Within the division of Geology and Geography, two committees on Paleoecology (1935-1937 and 1941) made a good start, however. The first of them included three zoologists: Paul Bartsch, Austin H. Clark, and M. W. de Laubenfels. In his first preliminary report the chairman, W. H. Twenhofel, said explicitly that the committee "was not organized to undertake research in the field of its study." Its object was "assemblage of data bearing on the ecologic or environmental relation of fossil organisms. . . ." The work of the committee was understood to be "in the nature of propaganda to bring the subject of paleoecology to the attention of students so that it may be talked about and become a part of the general thinking." This sort of service is as far-reaching as its effects are hard to trace. It may be said to be the first function of every organization that undertakes to stimulate research.

The work of this committee recalls that of the Committee on Submarine Configuration and Oceanic Circulation. But unlike the latter, it terminated its own existence after issuing but two reports (1936, 64 pages; 1937, 63 pages). Of individual groups, it had treated of vertebrates, arthropods in general and trilobites in particular, Paleozoic cephalopods, brachiopods, Porifera, and Foraminifera. Of environments, the Lower Part of Florida (with special emphasis on mollusks) was the only one covered. The Committee had obviously barely scratched the surface of this vast subject. Yet the student of sediments, as well as the paleontologist, urgently needs an understanding along the lines of ecology. Why was there no outcry of protest from workers in these fields when this valuable series stopped abruptly? Did they know about it at all? The division of Geology and Geography is taking the latter question seriously, as applied to all its products. There is good evidence that our division has been hiding its light under the proverbial bushel. Steps have been taken to make more widely known the first report of the new "Committee on Ecology of Marine Organisms" (Annual Report 1940-1941, Appendix A-1, 52 pages).

The new name of the second Committee reflects a change in the outlook of its personnel, which comprises chiefly men interested in the physiologic-oceanographic aspects of ecology, in the probing into the factors that control the character and distribution of life. The morphologic-zoogeographic-paleontologic element which prevailed in the first committee might well form a complementary group devoted to the presenting of the results produced by physico-chemical factors of the environment within the framework of space and time.

It is to be hoped that the present committee will eventually take the step which so greatly vitalized the Committee on Submarine Configuration and Oceanic Circulation. There is, of course, no "geobiological" counterpart to the Geophysical Union, but the end can be achieved by bringing together two types of men: one, geologists and paleontologists who recognize definite groups of facts which will be understood only when certain gaps in the knowledge concerning modern organisms have been filled; and the other, the biologists and oceanographers, who have the techniques and the interests required to fill those gaps. Out of this meeting of minds comes deliberate planning and directed research and from it that awakening of mutual interest in and respect for the other's field which is necessary for a deepening of understanding of problems and realities on both sides.[3]

But ecology is only one aspect of biology. Is there no need for getting together in other fields? "Evolution" is still full of riddles. The disciples of the young and vigorous experimental science of genetics have little time for, and less interest in, the body of solid fact which the older, undaunted historical disciplines of morphology, taxonomy, and paleontology are accumulating. Insofar as the disciples of the latter exhibit interest in genetics, it is mingled with distrust in the long-range significance of the former's findings. Here is a field that requires joint counseling. The new and independently created informal "Society for the Study of Speciation,"[4] the "dynamics of the origin of species," will go far in this direction. "The general object of the Society is to institute an informal information service which will tend to correlate the various approaches." The Paleontological Society was quick to invite its members to join. It is represented on the Executive Committee by Dr. G. G. Simpson, a leader among the vertebrate paleontologists who individually and as a group have kept most directly in touch with modern developments in biology. May we look to them to take the initiative in bringing together their own men with invertebrate and plant paleontologists for an enlivening discussion of "the dynamics of the origin of species," perhaps under the auspices of the National Research Council?

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[3] Similar developments are to be hoped for in the other fields of ecology, those of fresh-water and land environments.

[4] For information address the Secretary, Alfred E Emerson, Department of Zoology, The University of Chicago, Chicago, Illinois.

And how about zoogeography, past and present? The hypothesis of the drifting of continents was conceived and developed by the meteorologist Wegener and supported dominantly by zoogeographical observations. His hypothesis must have aroused interest in this fascinating field of knowledge among geologists and paleontologists. Paleobotanists have long cultivated energetically the field of phytogeography. Zoology, botany, and geology would benefit from activities that would open up the newest advances as well as the basic principles of zoogeography and phytogeography to a wider audience.

Even the lowly field of "ichnology," the study of tracks, trails, and burrows left behind by organisms, holds important keys to geological problems and deserves to be brought to more general attention. It has recently demonstrated the presence of King crabs (limulids) in the Triassic of New Jersey (Caster, 1939) and has disproved what for three-quarters of a century had been interpreted as evidence of frequent exposure to the air of the lime oozes of the Upper Jurassic embayment of the classic Solnhofen region (Caster, 1940). It should increase our understanding of pre-Cambrian life materially, if an interested group of workers made an intensive search for tracks and trails of what college texts still call "worms." The "very numerous" tracks of mammals, including small elephant and modern-looking canid, horse, and artiodactyl tracks, probably of Pliocene age, which Curry discovered on the east side of Death Valley (Noble, 1941, p. 956), are of great interest beyond paleontology. They may represent a relatively recent counterpart to the famous dinosaur tracks. If studied by men familiar with techniques of sedimentation, climatology, geomorphology, and mammal biology, they may help solve more precisely the question of the climatologic and topographic environment during the Triassic in the region of the present Connecticut Valley.

GEOLOGY PROPER

GENERAL STATEMENT

Turning from the borderline fields to geology proper, we recognize the same trend of development, from first efforts to create bibliographic tools and spread knowledge of, and interest in new or neglected lines of study, to a critical synthesis of knowledge that leads to co-ordinated investigations focused by the newly gained perspective. It is convenient to report the examples of these activities under corresponding headings.

BIBLIOGRAPHIC TOOLS AND REPORTS OF PROGRESS

Among the earliest publications of the Division of Geology and Geography two bibliographic lists stand out: E. B. Mathews' (1923) *Catalogue of published bibliographies in geology, 1896-1920* and H. P. Little's (1922)

List of manuscript bibliographies in geology and geography. Bibliographies or annotated lists of papers on specific subjects were contributed in separate issues by a number of committees as an effective way to open up a new or newly active field to geologists at large, as, for instance, isostasy (Adolph Knopf, 1924), the mechanics of igneous intrusion (F. F. Grout, 1935), and structural petrology (T. S. Lovering, 1938).

A contribution of major proportions, envisaged in one form or other since the earliest days of the National Research Council, became reality when under the leadership of Waldemar Lindgren a committee succeeded in raising a total of $21,358.58 from mining and oil companies and individual donors and thereby made possible the publication of the *Annotated bibliography of Economic Geology.* The first volume, for the year 1928, appeared in 1929, under the sponsorship of the National Research Council. In 1936 the sponsorship was transferred to the Society of Economic Geologists. For several years now the Geological Society has supported it liberally.

While speaking of bibliographies we may well point with pride to the *Bibliography and index of geology exclusive of North America.* With this publication added to the U. S. Geological Survey's *Bibliography of North American geology,* the first and only international bibliography of geology has been created in the English tongue. It has been made possible through the close and harmonious co-operation between the U. S. Geological Survey and the Geological Society.

Two committees have set unique records: the Committees on Paleobotany and on Sedimentation. The Committee on Paleobotany was among the first to be created in the Division of Geology and Geography (1919). Beginning in 1921,[5] it has reported annually, with the exception of an early gap of several years, the important papers published in North America during the year, together with comments on work in progress. Bound together, these reports represent a historical record of real value. Relatively few geologists and even paleontologists know about it. It deserves thoughtful consideration. Would not every unit in the large field of geology benefit by such an annual summary? Suppose we had a similar annual record of work accomplished, let us say on the structure of the Appalachians or the Great Basin, or on the mechanics of solids in its application to structural geology, or in petrology. Such a record would be no more difficult to create and would be no less valuable. In fact, if sufficiently condensed and cast into the right form, such annual reports might enable many workers in different fields to keep in touch with the essentials of progress along diverse lines of geology too far removed from their own. Such a work, clearly conceived and effectively carried out, would be an excellent antidote to the constriction of vision which accompanies all specialization.[6]

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[5] The first report appeared in 1923 But, in addition to the report on the current year it contained summaries for 1921 and 1922 by David White and E W Berry respectively.

Perhaps the most remarkable record has been set by the Committee on Sedimentation which, since its inception, has pursued vigorously a number of lines of usefulness. It has developed the policy of publishing progress reports (with bibliographies) on a given subject at unspecified intervals of a number of years. Each annual report contains several such progress reports. In volume, its annual reports[7] rose quickly from 5 pages in 1920 to 68 pages in 1923 and have averaged close to 100 pages since then. In addition to reports of progress, they contain papers concerned with the standardization of nomenclature and laboratory methods, and lately especially also summaries of new knowledge in allied fields such as hydrology and colloid chemistry. To secure greater completeness and precision in the description of sedimentary rocks, the committee prepared and had published a *Schedule for field description of sedimentary rocks* (M. I. Goldman and D. F. Hewett, 1923) and a *Color chart for the description of sedimentary rocks* (M. I. Goldman and H. E. Merwin, 1928). Two books have carried the direct influence of the work of this vigorous committee into every geological library and classroom in the English-speaking world: a symposium welded into a coherent work by W. H. Twenhofel in *Treatise on sedimentation* (1st edition, 1926; 2nd revised edition, 1932) and the symposium edited by Parker D. Trask in *Recent marine sediments* (Am. Assn. Petrol. Geol., 1939). These exceptionally influential works, like the committee's Annual Reports, contain in a large measure the sort of synthesis that creates perspective and reveals gaps in our knowledge. Work of that character is treated more specifically under the next heading.

CRITICAL SYNTHESIS OF DATA, ANALYSIS OF PROBLEMS, AND ORIGINAL INVESTIGATIONS

When the newly created Committee on Tectonics held its first meeting in December 1922, it voted "to commence work on a tectonic map of North America which should bring out the trends of folding, the principal lines of faulting, the axes of doming and related structural features." The immediate impulse for this ambitious undertaking had come from the epochal structure map of Eurasia which Argand had exhibited that summer at the meeting of the International Geological Congress at Brussels.[8] It could not be said to have grown from a specific desire on the part of American geologists for such a map. In fact, the word "tectonic map" did not convey much, nor the same idea, to most men. So the plan went through a period of incubation. "The time was not ripe," *i. e.*, we geologists were not ripe for this undertaking. When plans were made for the International Geological Congress to meet at Washington, D. C., G. R. Mansfield encouraged Philip B. King of the U. S. Geological Survey to prepare a general description of major structural units in the United States, accompanied by a small-scale tectonic map (published as Guidebook 28, 1932). A little later, Ralph D. Reed, in co-operation with J. S. Hollister, undertook the preparation of a tectonic map of Southern California which set a new standard and attracted wide attention (Reed and Hollister, 1936). These two publications represent high points on a rising wave of general interest. Under its impact, in 1935, under the leadership of Chester R. Longwell, the construction of a tectonic map was begun in earnest by the Committee on Tectonics. It was restricted to the United States and planned on a larger scale—that of the new geologic map of the United States. The compilation was undertaken by eleven subcommittees (Nat. Res. Council, Div. Geol. Geogr., Ann. Report for 1934–1935, Appendix D). Three years later, with the financial help of the American Association of Petroleum Geologists and The Geological Society of America, the maps furnished by the subcommittees were assembled in a first tentative draft. Two hundred black and white lithoprinted copies of this map were distributed among organizations and individuals for corrections and additions. The response was gratifying. The effective co-operation of the American Association of Petroleum Geologists and allied local societies deserves special praise. The final revision and drafting of the map is now in the able hands of P. B. King and will be completed as quickly as the present emergency permits. Arrangements for publication have been made with the American Association of Petroleum Geologists, one of three organizations that had expressed interest in the publishing of the map.

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[6] *Compare* Geologische Jahresberichte, begun 1938. (Gebr. Borntraeger, Berlin.)

[7] It is unfortunate that these reports were not published in uniform format and print. Reports 1 to 3 (1920-1922) were issued mimeographed and unbound; 4 to 8 (1922-1928) paper-bound mimeographed; 9 to 11 (1928-1930) printed in the Reprint and Circular Series as Nos 85, 92, 98; Nos 12 and 13 (1930-1934) printed as Bulletins 89 and 98 respectively; Nos. 14-18 (1936-1940) were again issued in paper-bound mimeographed form.

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[8] While speaking of the results of co-operation in geology we do well to recognise that we also owe to the catalytic action of the International Geologic Congress our two most important summary maps: (1) The creation of a geological map of North America (1:5,000,000) was proposed by J C Russell at the Ottawa meeting of The Geological Society of America (Dec 1905) in connection with the impending meeting of the International Congress at Mexico City in 1906 (G S A , Bull , vol. 17, 1906, p 709; Cong. Geol. Int. 1906, C.R , 1907, p. 211) It was produced, apparently in record time, under Bailey Willis' direction through the co-operation of the Surveys of Canada, the United States, and Mexico. A preliminary edition was published in 1906 at the expense of the Congress (U.S.G.S., 27th Ann. Report, 1906, p 21); (2) In that same year (1906) plans concerning a geologic map of the United States on the scale 1:2,500,000 are mentioned in the Annual Report of the U. S. Geological Survey For many parts of the country, however, not enough detail was available at that time for a map on so large a scale. Twenty years later, when the time was drawing near again for an invitation to the United States—in the winter of 1926-1927—following the meeting of the International Congress in Spain, the compilation of such a map was begun in earnest (U. S Geol. Survey, 48th Ann Report, 1927, p. 71). It was ready for distribution when the Congress met in Washington in 1933.

As in other cases, development of this project is sketched here in some detail because it illustrates well a number of matters that concern everyone interested in the development of a research program.

The story of the glacial map of North America is equally instructive. A committee was organized by the division in October 1939, under the leadership of R. F. Flint. In contrast to the tectonic map, the time was overripe for this kind of a map. There was no period of incubation. Work progressed rapidly. Gaps were filled by field work, some financed by the Geological Society. A symposium was held at the Boston meeting of the Society in 1941. The National Research Council division allotted a total of $1000 for the traveling expenses of committee members in order to facilitate the working out of the details for the first draft. The American Geographical Society contributed $1000 to secure the services of an experienced draftswoman for D. A. Nichols of the Geological Survey of Canada, who is giving practically his full time to the work of this committee. From the start, the Canadian Survey has set a fine example of official co-operation. The map is being drafted on a new base which is in process of being computed and plotted. The map will contain a great deal of new and unpublished information derived from the recent exploratory work of the Canadian Survey and numerous other sources.

The glacial map of North America will be published on the scale of 60 miles to the inch. It will be provided with submarine contours from data furnished for Canada by Nichols, for the United States by the Coast and Geodetic Survey through the interest of Lieutenant Paul A. Smith. The Geological Society is looking forward to publishing it.

These two maps, tectonic and glacial, when completed will be a challenge to American geologists. They will be criticized for some of the things shown on them and more for what is not shown. But, above all, they will be used and they will lead to new thought, new work, new knowledge, all on a larger perspective than has been possible without such attempts at synthesis.

The new correlation charts of the Committee on Stratigraphy perform a similar service. This committee does not deal with the shape and structure of sharply circumscribed bodies, but with the relative position on the time scale of larger and less easily definable categories, the formations. Both, the definition of these units and their relative incidence in time, are in many cases matters of judgment rather than of direct observation. It is, therefore, difficult to achieve agreement among different workers. Although begun in 1932, this work has not been completed yet.

Two of the twelve charts and a general introductory statement have appeared in the March and April issues of the Society's Bulletin for 1942; five others are virtually ready for the printer, and work on the remaining five is in progress.

A fine example of a sharply focused, broadly co-operative enterprise of a different kind is that which finally resulted in the publication of the Society's Special Paper No. 24, Dec., 1939: *Contributions to a knowledge of the lead and zinc deposits of the Mississippi Valley region.* This work was produced under the leadership of E. S. Bastin and edited by him. It began in 1932 with the formation of a subcommittee of the Committee on Ore Deposition. In the first two years, the committee met three times—at St. Louis, Mo., Chicago, Ill., and Baxter Springs, Kansas. A total of 45 people attended the last meeting, at which papers were read and excursions held. Graduate students at four universities and most of the members of the committee undertook special mineralogic, petrographic, and chemical studies. Stratigraphic and structural data were collected. At the end of five years, when the material was practically ready for the press, the committee asked to be discontinued, having secured a noteworthy addition of knowledge to a basic problem of ore genesis. (*See* Annual Reports of the Division from 1933 to 1937)

In 1936, when W. H. Newhouse became chairman, the Committee on Processes of Ore Deposition took steps to prepare a symposium on the influence of structure in localizing mineralization. Over 70 manuscripts have been edited and combined in a large volume which is being published by the Princeton University Press.

It is, of course, impossible to speak here of the work of all the committees of the Division, through the last 20 years. Two further examples illustrate still different ways of securing results. Grout's Committee on Batholith Problems, which functioned between 1928 and 1935, issued two thought-provoking reports: *Problems of batholiths* (rev. ed., 1933) and *Comments on magmatic stoping* (1935), the former a broad analysis, the latter chiefly a collection of quotations.

The work of the Committee on Shoreline Investigations has produced two unique results. Studies on shore phenomena, in co-operation with the U. S. Coast and Geodetic Survey, were among the first suggestions to be made in 1919 to the Division of Geology and Geography of the newly created National Research Council. In 1923 a Committee on Shoreline Investigations was appointed with Douglas Johnson as chairman. This committee undertook two projects: first, a specific one, *viz.*—to test the theory of local variations in mean sea level. The New York waters were chosen for the test. The U. S. Coast and Geodetic Survey, the Department of Docks of New York City, and the Department of

Plant and Structure of New York City co-operated with the National Research Council. The results of this investigation were published by Douglas Johnson in Bulletin 70 of the National Research Council (1929).

The other project was a very broad one: to stimulate interest in the systematic collection of data concerning changes in the beaches of the Atlantic seaboard which might later be studied "collectively, to deduce the underlying principles, and separately, to ascertain the variations from these principles to which individual beaches are subject." (Ann. Report, 1927, App. F, p. 3.) In order to create interest in this subject, with its wide social and economic implications, a circular letter was sent in 1926 to some two hundred addresses along the Atlantic and Gulf coasts, stating the Committee's

> "conception of the problem, making certain specific inquiries regarding conditions along the different sections of the coast, and requesting an expression of opinion as to the feasibility of securing public support for the cooperative study which the committee had in mind."

Ensuing correspondence, contact with officials of New Jersey, and resulting conferences led eventually to the formation, in the same year, of the American Shore and Beach Preservation Association. This active organization, in turn, has been a factor leading to the formation of the Beach Erosion Board in the Corps of Engineers of the United States Army. Although not directly connected with the Research Council, the American Shore and Beach Preservation Association usually holds its annual winter meeting in the National Academy-Research Council Building. In 1933 it started publication of a new journal "Shore and Beach." The work of the Division's committee which was discontinued in 1935 is thus being carried on on a broad basis.

OUTLOOK

Along purely geological lines no less than in the borderlands large problems invite joint planning and action. Let us round out this discussion by examining a few.

Regional studies, especially of major structural units, suggest themselves at once. In fact, the systematic study of the Rocky Mountain Region was one of the very first projects suggested to the newly formed National Research Council (Blackwelder, May 1, 1919). Such studies have been organized successfully by members of The Geological Society of America with the support of university departments, as, *e. g.*, Kellum and Imlay's studies in the folded belt of northern Mexico[9] and the work of the Yellowstone-Bighorn Research Association.[10] Neither of these was undertaken deliberately as part of a program by the Geological Society as such, nor were they financed exclusively by it, though both received liberal assistance.[11]

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[9] This has resulted so far in the publication of one book (by the University of Michigan), about 10 papers that have appeared in the Bulletin of The Geological Society of America, and one printed by the American Association of Petroleum Geologists.

Where important gaps exist in our knowledge of critical regions, planning will yield important results within our own borders and, more so even, in the vast regions of Latin America. The obstacle that blocks attempts at larger planning in such directions is a curious, indiscriminate fear of areal mapping, as if regional structures could ever be grasped without it. No one would propose random mapping for mapping's sake, nor is it our function to extend further any survey into territories that may not be expected to shed new light on some larger question. But we cannot afford to frown on an investigation because it involves areal mapping as an essential approach to a vital problem of structure.

There are, of course, large regional problems which require other methods for their solution. One concerns the synchronism and the extent of mid-Pleistocene continental deformation. It has been suggested by the writer that most of the United States and perhaps this and other continents as well have experienced a major epeirogenic uplift during mid-Pleistocene time which was responsible for the rejuvenation ("deep valley" stage) of stream valleys in the mid-continent regions as well as in those of the Cordilleran Mountains and which may have been contemporaneous with the mid-Pleistocene folding of the Californian Coast Ranges (Bucher, 1940, p. 422, footnote). This is a large problem that can be solved only by geologic methods and actually goes as deep into the problem of crustal deformation as any investigations by geophysical methods can hope to do. It is the sort of problem that should be attacked in an organized, co-ordinated effort. Here is material for a future major project of the National Research Council or the Geological Society or both.

Closely allied with it is another question of continent-wide interest. From a cross section of the terraces in the Susquehanna River valley, M. R. Campbell has attempted to compute the relative volumes of rock that were removed from the valley in the intervals between successive uplifts (Campbell, 1933, p. 572). Would it not be of interest to know what these ratios are for typical streams all over the country? Here is another project calling for co-operative, country-wide study from which valuable insight into time relations of epeirogenic processes will result.

The last project serves to emphasize the general need of quantitative data concerning geological processes We need a companion volume to

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[10] This has led so far to the publication of about 85 scientific papers in different periodicals

[11] Kellum's work also received a grant-in-aid from the National Research Council

Birch's *Handbook of physical constants,* one giving reliable quantitative data concerning rates involved in all sorts of geological processes and the factors that control them.[12] This leads straight to the borderland between geology and applied physics such as soil erosion, hydrology of running water and shore lines, and meteorology.

The manner in which the stratigraphic timing of strata ("correlation") is recorded is highly unsatisfactory. The writer has discussed this matter in a paper given at the Boston meeting entitled *Method proposed to introduce the concept of "limits of error" into the stratigraphic timing of tectonic movements.*[13] He believes that improvements along the lines suggested by him should be attempted. This can best be done by a concerted joint effort.

A unique opportunity for studies bearing on the major problems of crustal deformation is offered by the photographic record of the Ross Shelf Ice which has been secured by Byrd's Antarctic Expeditions. Locally, as, *e. g.,* in the Bay of Whales, the ice is thrown into systems of folds or is traversed by large rifts or cut by systems of fractures (Gould, 1935). Much of the ice rests on a rock shelf, some of it floats. Most of it is in movement. A committee of the Division of Geology and Geography of the National Research Council, just organized, has undertaken to prepare a tectonic map of the Ross Shelf Ice and to attempt an analysis of the mechanics that determine the nature and distribution of the structures. It is hoped that this huge laboratory experiment of nature will yield new clues for an understanding of the deformation of the earth's crust.

Fringing the borders of the continents, the marine terraces offer a problem parallel to that of the river terraces mentioned above and directly related to it. About 10 years ago, Douglas Johnson submitted a project to the Geological Society which provided for a series of co-ordinated, simultaneous studies of typical sequences of terraces on different shores along the Atlantic Ocean. Like another broadly conceived plan that was submitted by A. C. Veatch about the same time, this project was shelved at the time for various reasons. But as our vision and experience grow and we learn to think in larger terms, we shall return to this kind of investigation. Others have pointed to the example of the American Geographical Society which with its superb Hispanic American mapping program has set a unique example of courage and vision in long-range planning.

It is not worth while to prolong this looking ahead in the field of geology. For a veritable treasure of "Suggestions concerning desirable lines of research in the field of geology and geography" the reader is referred once more to the report of that title which was issued by the Division in 1906 under the editorship of E. S. Bastin for physical geology, C. O. Dunbar for Paleontology and Stratigraphy, and R. S. Platt for Geography.

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[12] *See, e.g.,* Fenneman's comments on *The actual rates at which rocks waste away* (Fenneman, 1938, p. 146).

[13] Publication deferred

One item barely touched upon in that report deserves being set forth in some detail. It will close this discussion fittingly.[14]

The modern methods of echo sounding have at last brought within our reach definite information concerning the geomorphology of the vast spaces covered by the ocean floors. Many echo-sounding profiles are accumulating in the files of the Coast and Geodetic Survey and the Hydrographic Office of the Navy. A year ago the Committee on Research of the Geological Society directed the Council's attention to the value of these records. It is necessary that steps be taken to utilize them before the impression gains ground that further additions of new material and even the storage of the old represent waste.

These profiles give a picture of the configuration of the sea floor, not only in two dimensions, but in three, provided a large number of parallel and intersecting profiles is used. Unfortunately they tell little at first sight, and they must be used with full appreciation of all possible sources of error. The mere publication of a few selected profiles or hasty generalizations by one not thoroughly familiar with their use would almost certainly do harm.

We need men who deliberately train themselves for a field that is almost wholly new. To be successful in this field a man must have a thorough grasp of the problems of modern tectonics and geomorphology. He must first gain experience in three-dimensional analysis of surface forms based on profiles of greatly exaggerated scale. The best results, for a long time to come, will be attained in submarine geomorphology from the analysis of the profiles themselves. But contouring—on maps of relatively small scale—will have its place. Both methods presuppose great care in securing the most precise location possible for all the data on maps.

This means that in addition to a thorough geologic-geomorphologic training, the novice in this field should become familiar with the methods of echo sounding and the problems of position finding. This can be achieved best by studying the methods and practice of the Coast and Geodetic Survey. That organization, in times of peace, is usually seeking at least one or two adequately prepared and capable young men for various phases of survey and cartographic work.

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[14] The Subcommittee on Ocean Bottoms of the Geological Society's Committee on Research Program was directed by the Council (October, 1941) to call the attention of the membership to the promise which the relatively new field of suboceanic geomorphology holds The subcommittee has decided to embody in this address the following statement prepared by the chairman of the subcommittee, Lieut. Paul A Smith, and the writer

Such work, by the way, would give the student an opportunity to get acquainted with an occupation which he may later choose as his life work, if he so desires. If, however, he prefers to stay in geology, he should have little difficulty in finding a place in the academic world, provided our universities keep their doors open to men whose minds travel new roads.

CONCLUSION

With this vision of a way to grasp the geomorphology of the larger part of our globe that lies concealed beneath the sea, this discussion of research in geology and the National Research Council comes to an end. The story of co-operation in research embraces all our geological organizations. There is no clear line of demarcation between a "research program" of The Geological Society of America and that of the National Research Council. Whether there shall be one or not does not concern us here. Whatever is done officially, the greatest possible freedom should be permitted for the formulation of projects to suit the whims, prejudices, and jealousies of individuals or groups to which science is an unwilling heir. If this manner of growth appears less orderly than one might wish, it is also a secret and great source of strength. It is the very essence of democracy, as long as we stand ready to welcome and support worthy work from whatever source it may spring.

POSTSCRIPT

The essence of this paper was given orally at the Boston meeting of the Society, only a few weeks after Pearl Harbor. Since then our outlook has been turned rapidly to the cruel needs of the hour. The words "research" and "co-operation" have taken on a harder meaning. The offices of the National Research Council are rapidly being geared into the nation's war work, and members of the Geological Society are leaving their posts in increasing numbers to go into active service of one kind or another. Already the preoccupation with problems of pure geology, pursued without thought of power or possession, begins to seem remote and perhaps almost futile.

Let us not be misled. No matter how long it may last, the war will be over soon when reckoned by generations. The material destruction can and will be repaired. It is imperative that we keep our minds turned firmly to that future which justifies today's sacrifices, when those of us who survive and those who will follow our steps will return to the laboratories and studies. To them the preceding pages are addressed above all, to be read when they again take up the threads of thought and resume the creative dreams of free inquiry into nature.

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FLOOD DEPOSITS OF ARROYO SECO, LOS ANGELES COUNTY, CALIFORNIA

BY W. C. KRUMBEIN

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ABSTRACT

This study represents a continuation of earlier work in San Gabriel Canyon. Deposits left in Arroyo Seco by the flood of March 2, 1938, were mapped in the field, and samples were taken to determine how sedimentary characteristics varied with distance from source. In an area of Lowe granodiorite near the headwaters of the stream size and roundness of particular pebbles, free from contamination, were studied As in San Gabriel Canyon, the flood deposits were arranged largely in log and boulder jam deposits, which formed a series of partially overlapping features along the valley. These jam deposits have reverse size gradients,—mean size of the boulders increased markedly downstream.

In general, however, neither the size nor the sphericity of pebbles in the flood deposits, collected in corresponding portions of successive log jams, showed any change along the valley; roundness, on the other hand, increased noticeably downstream. These results agree with findings in San Gabriel Canyon and suggest that size and shape respond to the dynamical conditions at the immediate site of deposition, whereas roundness is a function of the abrasion suffered by the particles over the distance traveled.

An attempt is made, in the light of available evidence, to reconstruct the physical conditions governing the movement and deposition of the flood material, as well as to evaluate the relative roles of wear and selective transportation during floods.

INTRODUCTION

This paper continues the study of deposits associated with the flood in the Los Angeles region on March 2, 1938. In an earlier paper (Krumbein, 1940) the flood deposits of San Gabriel Canyon were described, based largely on a laboratory study of selected samples. This later study was done during the summer of 1940.

In many of the canyons the flood deposits either were not adequately exposed or they had been removed in part by road building and other operations. Conditions in Arroyo Seco, however, were almost ideal. This canyon had been closed to the public since the flood, and in June 1940 the deposits were still fairly continuous and well exposed. Six weeks were spent in a detailed study of parts of the canyon. Field evidence was supplemented by laboratory studies on samples, as well as by data obtained from engineers and others who had studied the effects of the flood.

Additional work was done in San Gabriel Canyon to complete details and test tentative conclusions. In general, the new work supports previous findings, although some modification of detail was necessary.

More time and a closer field examination of the deposits in another canyon affected by the same flood permitted the further testing of some tentative conclusions of the San Gabriel Canyon study and a better classification of flood deposits. The experience gained permits at least qualitative conclusions regarding the mechanics of debris movement and deposition by floods, as well as a statement of the problem of wear and selective transport along the canyon.

ACKNOWLEDGMENTS

This project was made possible by a grant from the Penrose Bequest of The Geological Society of America. The writer also wishes to acknowledge his indebtedness to numerous individuals who aided in the work. He is particularly indebted to Mr. Lyman Huff, a graduate student at the University of Chicago, for able assistance in the field and laboratory throughout the summer. Dr. E. Wayne Galliher (Barnsdall Oil Company), Dr. Ian Campbell (California Institute of Technology), Dr. James Gilluly (University of California at Los Angeles), Dr. Vito Vanoni (U. S. Soil Conservation Service, Pasadena), and Mr. D. M. Ilch (U. S. Forest Service, Glendora) were particularly helpful in their suggestions and criticisms of the work. Mr. M. F. Burke of the Los Angeles County Flood Control furnished much valuable information regarding the engineering problems associated with the flood. The Department of Geology of the California Institute of Technology kindly furnished office facilities, and the Soil Conservation Service extended the privileges of its Fineness Laboratory for mechanical analyses.

GEOLOGICAL SETTING OF ARROYO SECO

Arroyo Seco rises in the San Gabriel Mountains and flows generally southwest and south to its junction with the Los Angeles River within the city limits (Fig. 1). The area studied extends from the headwaters to the mouth of the canyon. The rectangle in the upper right of Figure 1 shows the area studied previously along the North Fork of San Gabriel River (Krumbein, 1940).

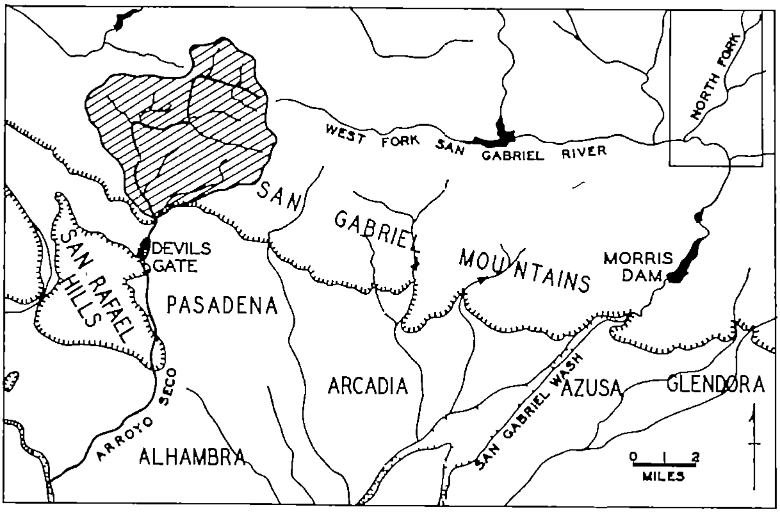


FIGURE 1.—*Index map of area studied*

The shaded area is the drainage basin of Arroyo Seco Canyon. The rectangle at the upper right shows the part of San Gabriel Canyon studied earlier

Practically the entire drainage basin of Arroyo Seco is shown on the Mt. Lowe (Los Angeles County) topographic map (1934). Figure 2 shows the extent of the basin and indicates all tributaries of Arroyo Seco within the mountain belt. The head of the main stream rises at Red Box Gap, at an elevation of about 4600 feet. The stream flows generally west and south for about 14 miles to the mouth of the canyon at the southern apex of the map where the elevation is 1150 feet. The average gradient is about 240 feet per mile. The canyon is steep-walled, with numerous rock exposures. Chaparral covers the gentler slopes, and the canyons themselves are lined with oaks and other trees. The numbers and letters on Figure 2 refer to sample localities, which will be described under appropriate heads.

Miller (1928; 1934) has described the geology of the San Gabriel Mountains. In general, the canyon cuts through a complex of crystalline rocks of possible pre-Cambrian and late Jurassic or early Cretaceous age. Several faults cut the region and locally control the trend of the main canyon. The mountains themselves are elevated fault blocks rather than folded belts. The present paper is not concerned directly with the bedrock geology of the region, except to the extent that it controlled the kinds of debris. The most important rock in the drainage basin is the Lowe granodiorite of possible late Jurassic or early Cretaceous age, which underlies much of the eastern part of the drainage basin. Figure 3 shows the areal distribution of this granodiorite, taken from Miller's map (1934). The main stream cuts this igneous mass in the northeastern part of its course, but beyond its downstream point of emergence from the granodiorite it flows entirely through rocks distinctly different from the Lowe granodiorite. Moreover, between the unnamed tributary which enters the main stream below the northern bulge of the granodiorite (locality D, Fig. 2) and Millard Canyon (locality R, Fig. 2) no other tributary taps the rock. Thus the Lowe granodiorite

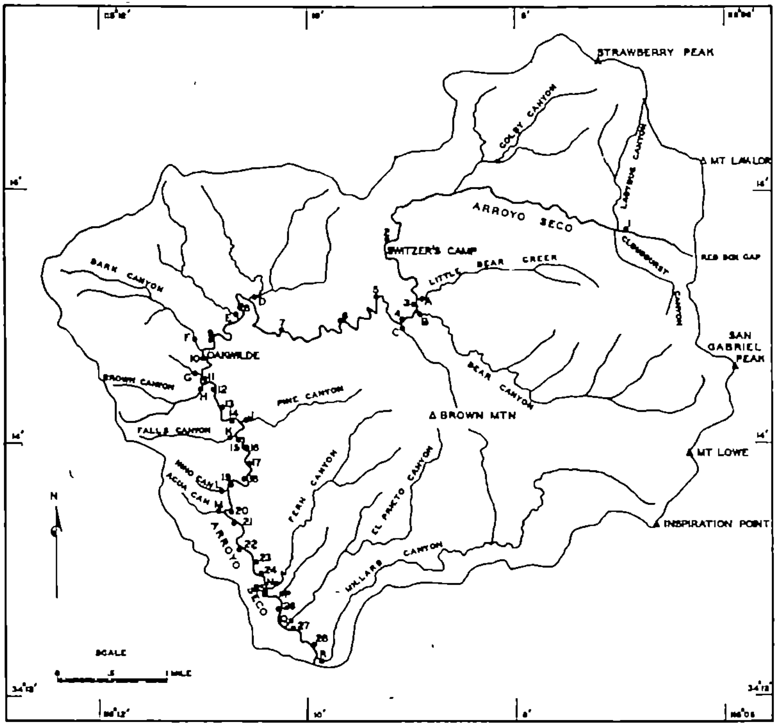


Figure 2.—*Arroyo Seco Canyon drainage basin*

can be used to study the progressive rounding of pebbles during transport, with no danger of contamination in the lower half of the stream. Millard Canyon itself enters the main stream at the latter's point of emergence from the mountains and so does not change the situation.

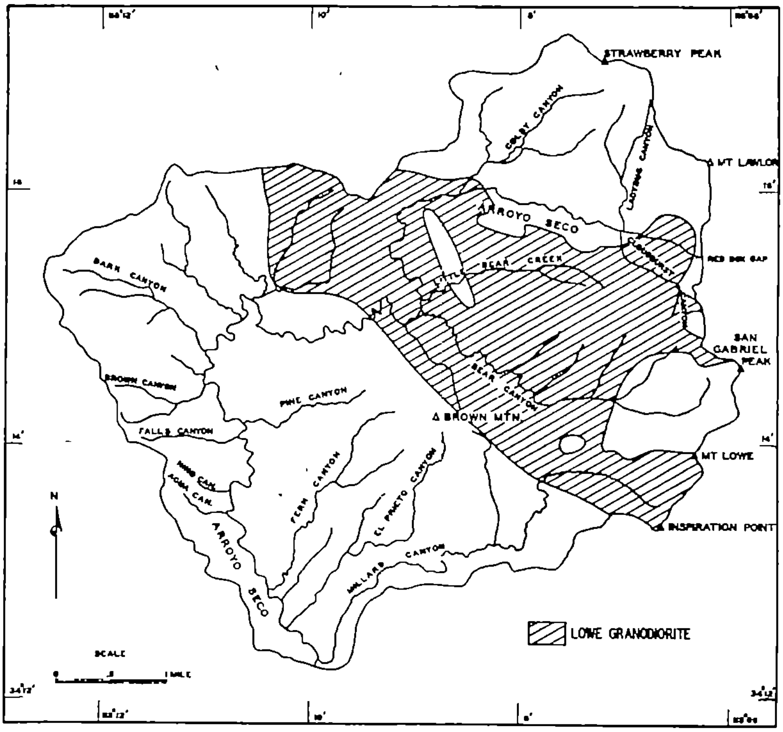


Figure 3.—*Distribution of Lowe granodiorite in Arroyo Seco*

FLOOD OF MARCH 2, 1938

GENERAL STATEMENT

The flood of March 2, 1938, equalled or exceeded any previous flood of record in Los Angeles County. The runoff peaks for the larger catchment areas equalled or exceeded any on record, and debris production was the greatest of record. Volumes of material eroded were estimated to vary from 12,000 to 80,000 cubic yards per square mile and averaged 51,000 cubic yards per square mile. This average is equivalent to the removal of more than a half-inch layer of debris from the entire area affected by the flood.

The Los Angeles Flood Control studied the engineering aspects of the flood (Burke, 1938), and through the kindness of Mr. Burke the following details are given. Mr. D. M. Ilch of the U. S. Forest Service made available information on the sources of the flood deposits in Arroyo Seco, as well as estimates of the flood damage.

CONDITIONS BEFORE THE FLOOD

The catchment basin of Arroyo Seco supplies water for Pasadena. Conservation practices are followed, and although an attempt is made to maintain a vegetation cover, in the past half century eight fires have swept parts of the basin. The most important recent fire, the Brown Mountain fire of 1934, burned an appreciable area in the center of the basin and made available a considerable amount of debris for movement during spring floods. Between 1934 and 1938 much silt had been carried down the tributaries into Arroyo Seco, and to a large extent this debris remained within the canyon area until the major flood. Within the last 10 years also, the Angeles Crest Highway was built through the mountainous area, involving numerous fills at steep canyon heads. During the 1938 flood this relatively loose material was avail-

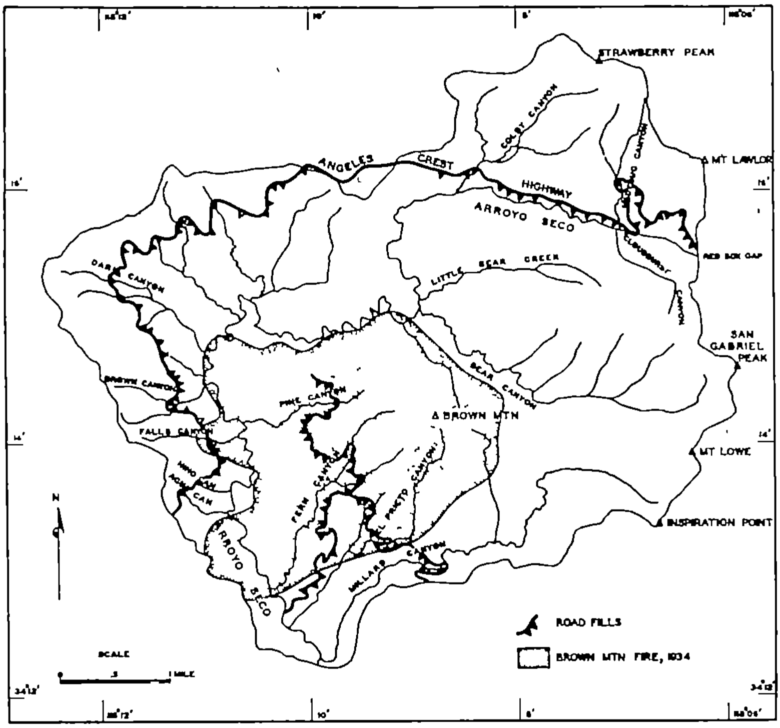


Figure 4.—*Sources of debris in Arroyo Seco flood of March 2, 1938*

able for transport by the flood waters. In addition, the steep mountain slopes were veneered with weathered debris which became unstable during the flood and caused numerous landslides. Fern Canyon, Pine Canyon, and Bear Canyon, as well as some parts of the main valley itself, lie within or adjacent to the area burned by the Brown Mountain fire of 1934 (Fig. 4).

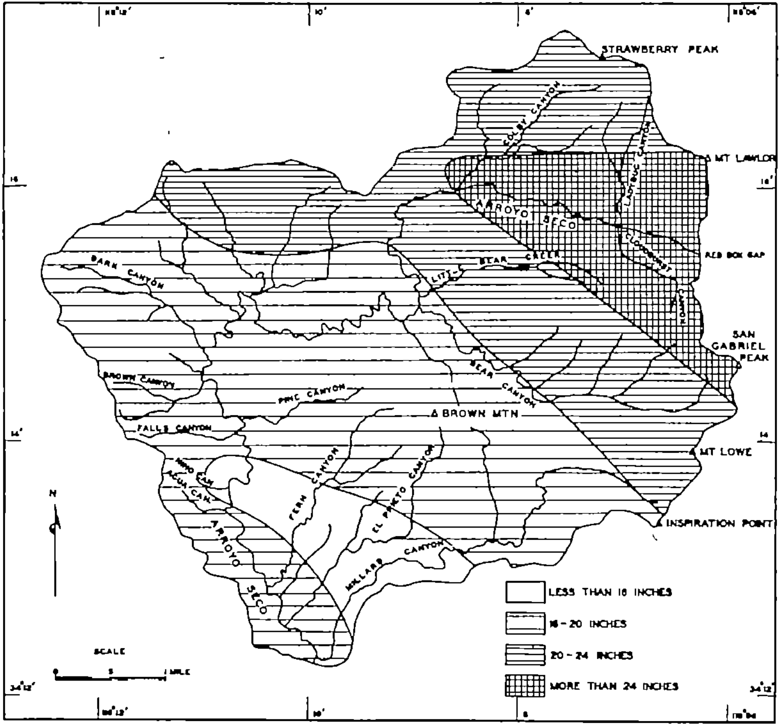


Figure 5.—*Generalized isohyetal map of Arroyo Seco*

Before the flood the Arroyo Seco drainage basin had considerable unconsolidated material. Normal spring floods moved part of this along steep tributaries to the main canyon, but the lessened slope in the main valley prevented transportation to the alluvial fan. The flood of March 2, 1938, however, caused a major shifting of debris along the main canyon.

PROGRESS OF THE FLOOD

Rainfall data.—The Los Angeles Flood Control compiled rainfall maps for the storm. Figure 5 is a generalized isohyetal map of the entire storm, from February 26 to March 3, 1938. In the Arroyo Seco catchment area the precipitation varied from less than 16 inches near the mouth of the canyon to more than 24 inches near the headwaters. During the first several days of the storm the rainfall was moderate, but on the evening of March 1 it increased abruptly. The moderate rainfall saturated the mantle rock, so that the more intense downfall resulted in excessive runoff. Generally speaking, the rainfall during the last period of the storm amounted to about 50 per cent of the total precipitation, which means rainfall intensities as high as 12 inches or more during a 24-hour period in some portions of the mountainous area.

The runoff associated with the flood varied during the storm. Previous to the intense downfall it is estimated at between 15 and 20 per cent. This left between 5 and 8 inches in the mountainous areas to saturate the mantle rock. During the second part of the storm the runoff increased to between 33 and 64 percent, the larger runoff associated with the larger drainage basins. In Arroyo Seco Canyon, a drainage area of 30.6 square miles, the runoff during the peak flood was 43 per cent. This excessive runoff resulted in very rapid accumulation of flood waters on March 2.

Stream discharge.—Estimates of maximum discharge along the canyon were made by engineers from high water marks. At Oakwilde the maximum was of the order of 6500 cubic feet per second. At the Ranger Station (near site 26, Fig. 2) it was about 9000 cubic feet per second, and at the entrance to Devils Gate Reservoir, just below the mouth of Millard Canyon, it was 11,400 cubic feet per second. The velocity of the flood water varied considerably along the canyon; maximum values in constrictions are believed to be of the order of 20 miles per hour.

Movement of debris.—Estimates of debris movement during any one flood are difficult to make, but fairly reliable figures are available for the Arroyo Seco drainage basin. About 1 mile downstream from the canyon mouth Devils Gate Reservoir served as a gauge on both the runoff and debris movement. Previous surveys of the reservoir showed that between 1920 and 1934 the reservoir received debris at the rate of 1920 cubic yards per square mile of watershed per year. The total estimated debris for the flood of March 2, 1938, was 70,000 cubic yards per square mile, which is about a 35-fold increase in the rate of erosion. During this one flood it is estimated that debris movement into or immediately upstream of Devils Gate Reservoir amounted to 1319 acre feet.

The maximum 24-hour inflow of water into the reservoir during the flood was 7420 acre feet. From this figure and the siltation of the reservoir Burke calculated that the volume of debris moved was 17.8 per cent of the maximum 24-hour inflow.

Flood damage.—Flood damage in Arroyo Seco Canyon was estimated at about $380,000, aside from the loss incurred in reduced reservoir capacity. This was distributed as follows: summer homes and camps, approximately $68,000; water diversion works, $28,000; road damage along Angeles Crest Highway, $140,000; and Arroyo Seco road and other minor roads, $144,000. More than 320,000 cubic yards of fill had to be replaced in embankments along 13 miles of Angeles Crest Highway after the flood.

DESCRIPTION OF FLOOD DEPOSITS

GENERAL CONSIDERATIONS

The flood deposits along Arroyo Seco Canyon were best developed in its lower course. Near the headwaters, especially upstream from Switzer's Camp (Fig. 2), numerous scours contained large boulders and local debris. Fresh bedrock was exposed in many parts of the stream channel. Downstream, especially below Oakwilde, the valley floor was covered with debris. Plate 1 shows the contrast between these two parts of the valley.

The Mt. Lowe topographic sheet (1934) shows a rather narrow flood plain in the lower third of Arroyo Seco Canyon traceable upstream from Oakwilde. Beyond this point the valley narrows markedly and enters what is locally known as "Marble Canyon" with precipitous walls and a sinuous course. Some of the tributaries in this gorge enter the main canyon over waterfalls. Where the flood plain appears aggradation becomes predominant as is indicated by the rarity of rock exposures in the channel floor downstream, except at narrows and sharp bends, where scour was locally severe. Upstream from the flood plain, however, bare spots on the canyon floor are not uncommon, and the deposits are generally more limited in volume and frequency.

In detail the flood deposits of the upper and lower canyon show a marked response to local conditions. Variations in valley width, and especially bends in the canyon, greatly affect deposition and scour. Where the valley widens a plain of sandy gravel may be found; at bends are boulder and log jams varying markedly in size and thickness. At tributaries along the canyon are alluvial cones, some of which cross the greater part of the valley floor.

LOG AND BOULDER JAMS

General statement.—Log jams in San Gabriel Canyon have been described previously (Krumbein, 1940), but their better preservation in Arroyo Seco and the more detailed study warrant this fuller consideration. In longitudinal section the jams are lenses or wedges thickening

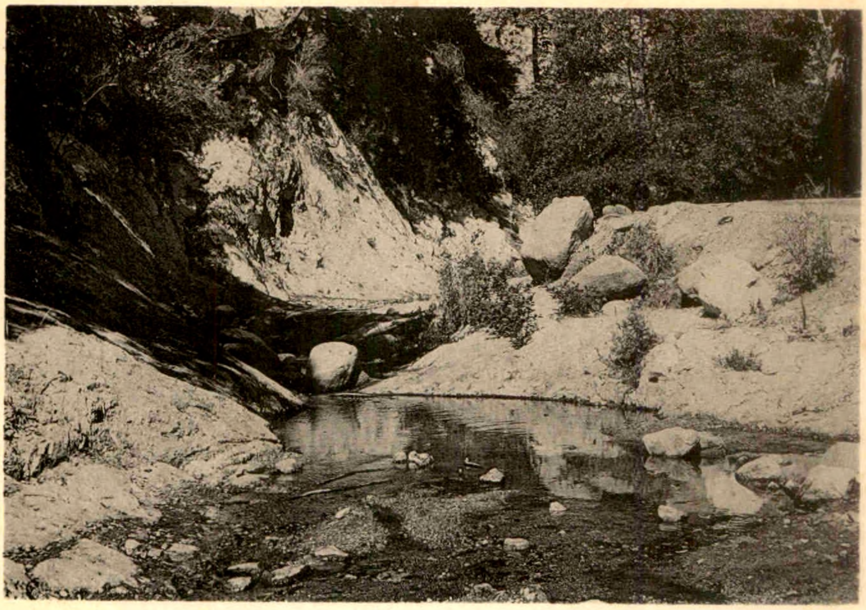


FIGURE 1. EXPOSED BEDROCK IN STREAM CHANNEL NEAR SWITZER'S CAMP

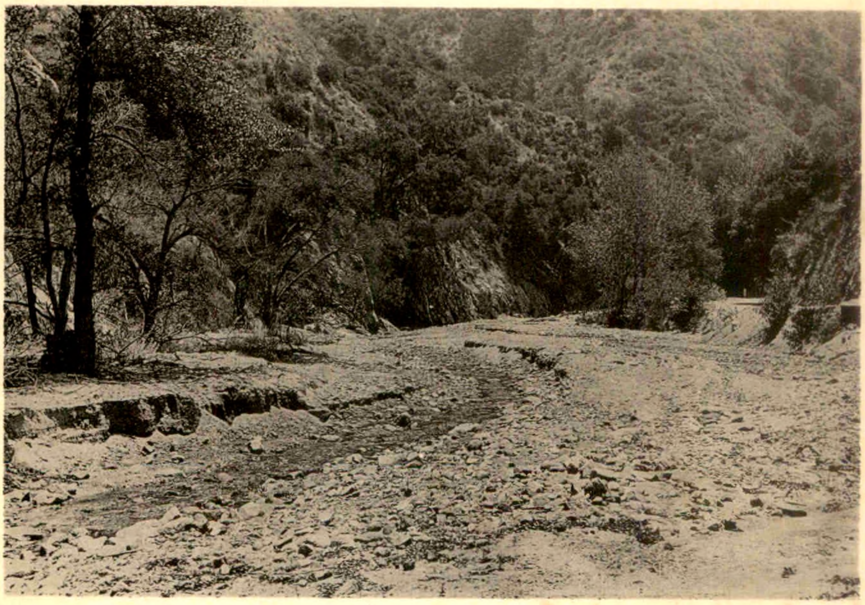
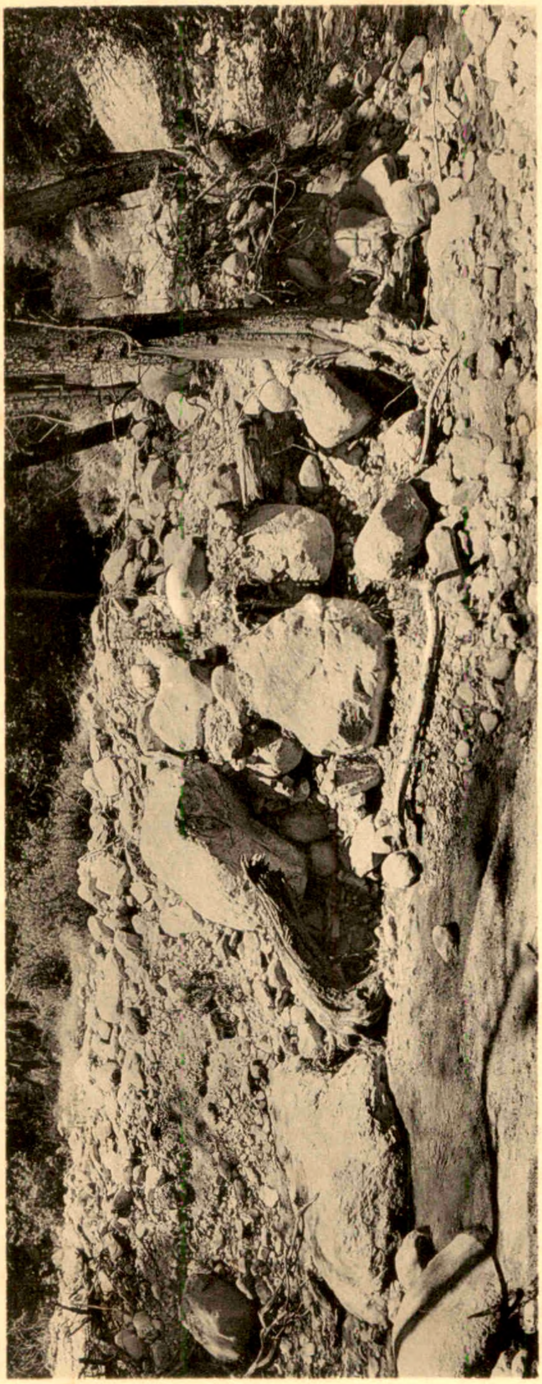


FIGURE 2. FLOOD DEPOSITS 1.5 MILES BELOW OAKWILDE

CONTRAST OF SCOURED AND FILLED PORTIONS OF ARROYO SECO



TERMINUS OF BOULDER JAM NEAR SITE 5

(Photo by Lyman Huff) Taken from canyon wall downstream from jam. Note bedded material at left and gradual increase of boulder size downstream to terminus. The original front slope of jam is preserved in the central portion of the picture, but some post-flood erosion occurred at the right. The present stream flows from right to left.

and coarsening downstream. The upper surface may lie essentially horizontal or slope gently downstream. In some jams the thickened downstream portion may locally rise above the general level.

Near the upstream end the material is frequently relatively fine, moderately well bedded, and fairly well sorted. Downstream the deposit thickens and coarsens, and bedding becomes more prominent. Near the downstream terminus appear numerous large boulders, both on the surface and within the deposit; the sorting becomes noticeably poorer; and the bedding appears to give way to a heterogeneous arrangement. The jam terminates abruptly in a relatively steep slope, which drops to an area of scour immediately downstream of the deposit. This scour area contains lag boulders and cobbles and merges into the next flood deposit still farther downstream.

Plate 2 shows the terminus of a boulder jam near site 5 in Arroyo Seco. The photograph was taken from the canyon wall along the scour area below the jam, and the view is diagonally upstream. The marked increase in boulder size is shown, as well as the heterogeneity of the boulders. On the left is the finer relatively well bedded material upstream. Generally this ideal arrangement is complicated by several stages of reworking. One jam may partially overlap another, or one deposit may be partially destroyed and a later accumulation may be formed at approximately the same location.

Figure 1 of Plate 3, shows the bouldery texture near the downstream terminus of a jam deposit, and Figure 2 of Plate 3 shows the size variation along a jam which has been cleared parallel to the valley. This figure demonstrates that the boulders actually are concentrated near the downstream end and are not merely buried by the upstream portion. Plate 4 shows the openwork seam at the base of the flood deposits and the degree of imbrication often observed in the central portions of the deposits. (The pencil in the figure points downstream.)

A well-preserved log and boulder jam was found at site 25 in Arroyo Seco (Fig. 6). The valley turns abruptly and narrows downstream. A bridge had been built at the turn, and the old road ran along the east side of the canyon upstream from the bridge. When the site was studied, the main part of the jam was well preserved, although there were evidences of shifting positions of the stream since the flood. Upstream (Fig. 6) the deposits were relatively fine sandy gravel with a surface of lag cobbles. To the south and east boulders increased noticeably in number and size, and at the terminus of the jam the front slope dropped abruptly some 15 feet to an area of scour downstream. Here the valley was swept clean, and all the fill adjacent to the bridge foundations had been removed. Fresh scours were visible on the exposed rocks along the valley wall, and except for large lag boulders this part of the valley was relatively free of debris. An idealized longitudinal section of the jam is shown in Figure 6. Numerous standing trees were associated with the deposit, but these are not shown.

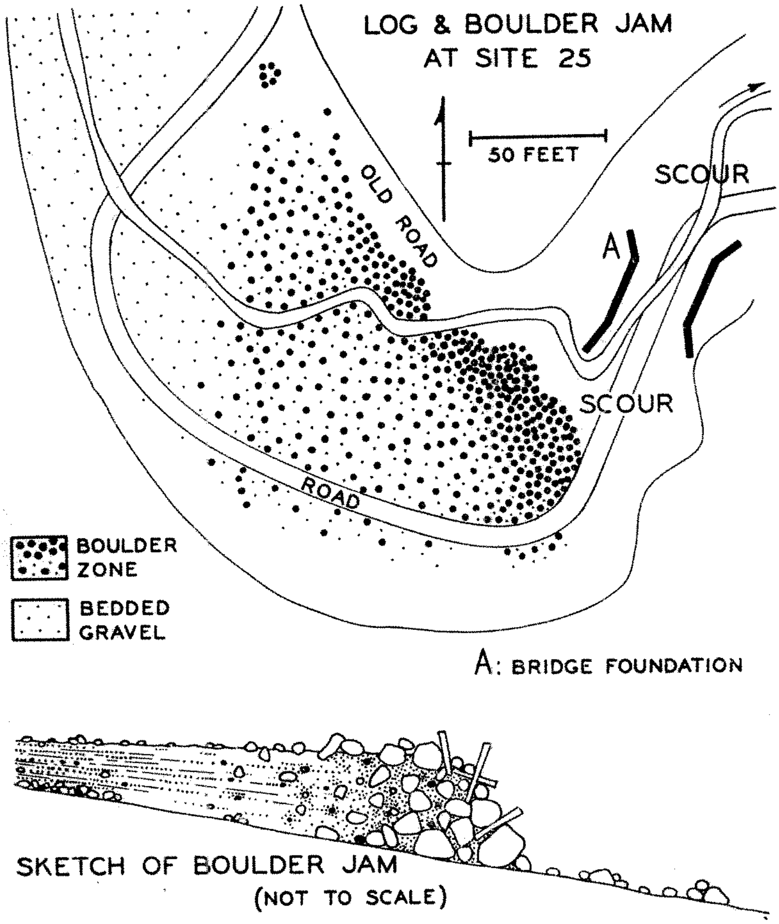


FIGURE 6.—*Map of boulder jam at site 25*
Lower figure is a sketch of the longitudinal section.

Size variation of boulders.—The striking changes in size of the jam material is illustrated by parts of a small jam near site 21 in Arroyo Seco (Pl. 5). The photographs were taken approximately 35 feet apart, and the adjacent strips were mounted close together to bring out the

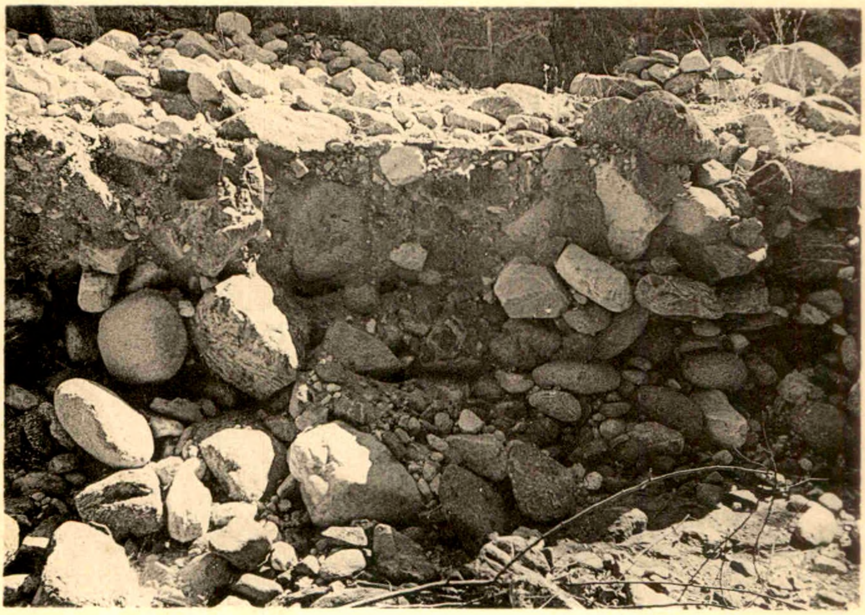


FIGURE 1. DOWNSTREAM END OF JAM NEAR SITE 8, SHOWING BOULDERY TEXTURE

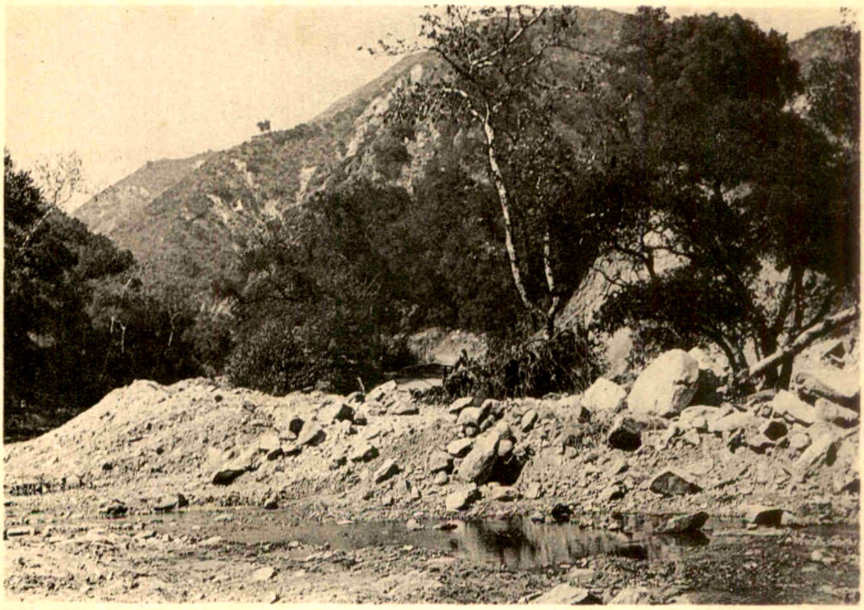


FIGURE 2. LONGITUDINAL SECTION ALONG JAM CLEARED PARALLEL TO SIDE OF VALLEY
This photo shows that the boulders are actually concentrated at the downstream end, and not merely buried by the upstream portions.

DETAILS OF JAM DEPOSITS

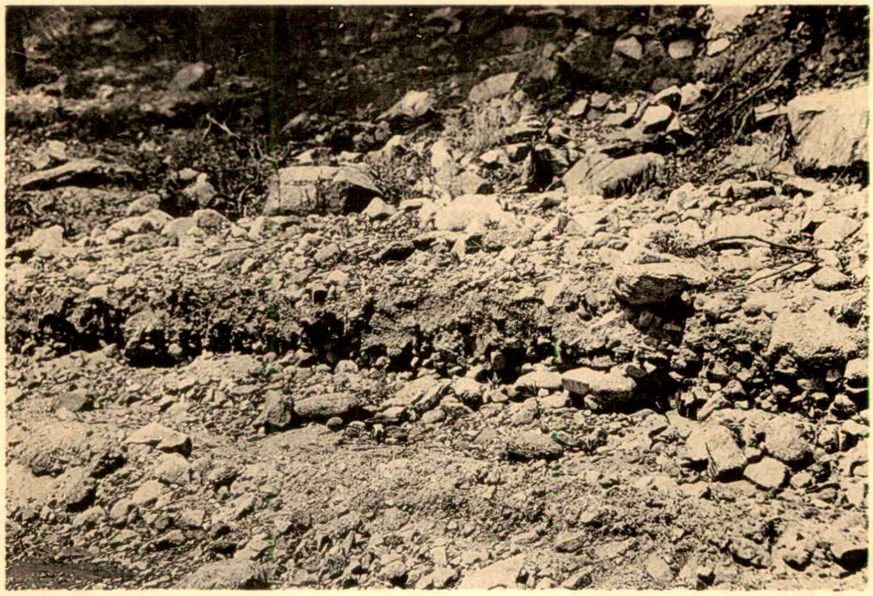


FIGURE 1. OPENWORK SEAM NEAR BASE OF DEPOSIT, SITE 14

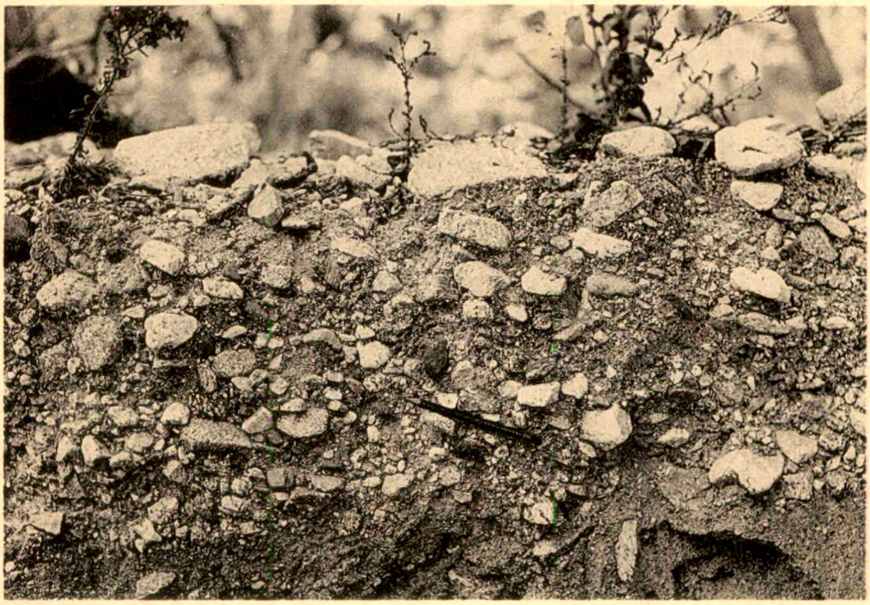


FIGURE 2. IMBRICATION OF PEBBLES IN CENTRAL PORTION OF DEPOSIT NEAR SITE 18
Pencil points downstream.

DETAILS OF JAM DEPOSITS

change in size. The downstream direction is downward in the plate. The largest boulder is about 4 feet in diameter.

The boulders in several jams were measured to estimate the magnitude of the size change. The material was too coarse for complete mechanical analyses, but for a first approximation sites were chosen at intervals of 25 feet along the jam, and the 20 largest boulders or cobbles at each site were measured. A knotted twine was held parallel to the intermediate diameter of each boulder, and this diameter was classified according to the Wentworth scale. The size classes ranged from 64 mm. (about 3 inches) to 2048 mm. (about 82 inches). The 20 values at each site were averaged to find the geometric mean boulder size (Fig. 7). In the jam near site 21 (Fig. 7, middle graph; *compare* Pl. 5), measured over a distance of 200 feet, the mean boulder size increased from 100 mm. to nearly 500 mm. In the jam at site 17 (Fig. 7, upper graph) the increase along 300 feet was from 200 mm. to 300 mm. At site 25 the slope

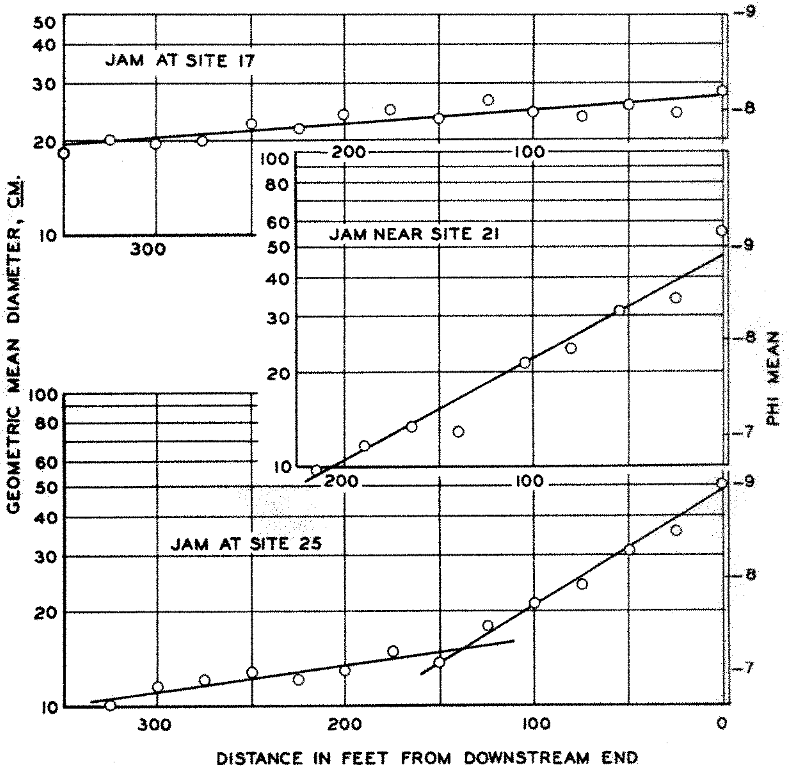


FIGURE 7.—*Size variation of boulders in jam deposits*

of the line (Fig. 7, lower graph) increases along the jam, suggesting a composite deposit.

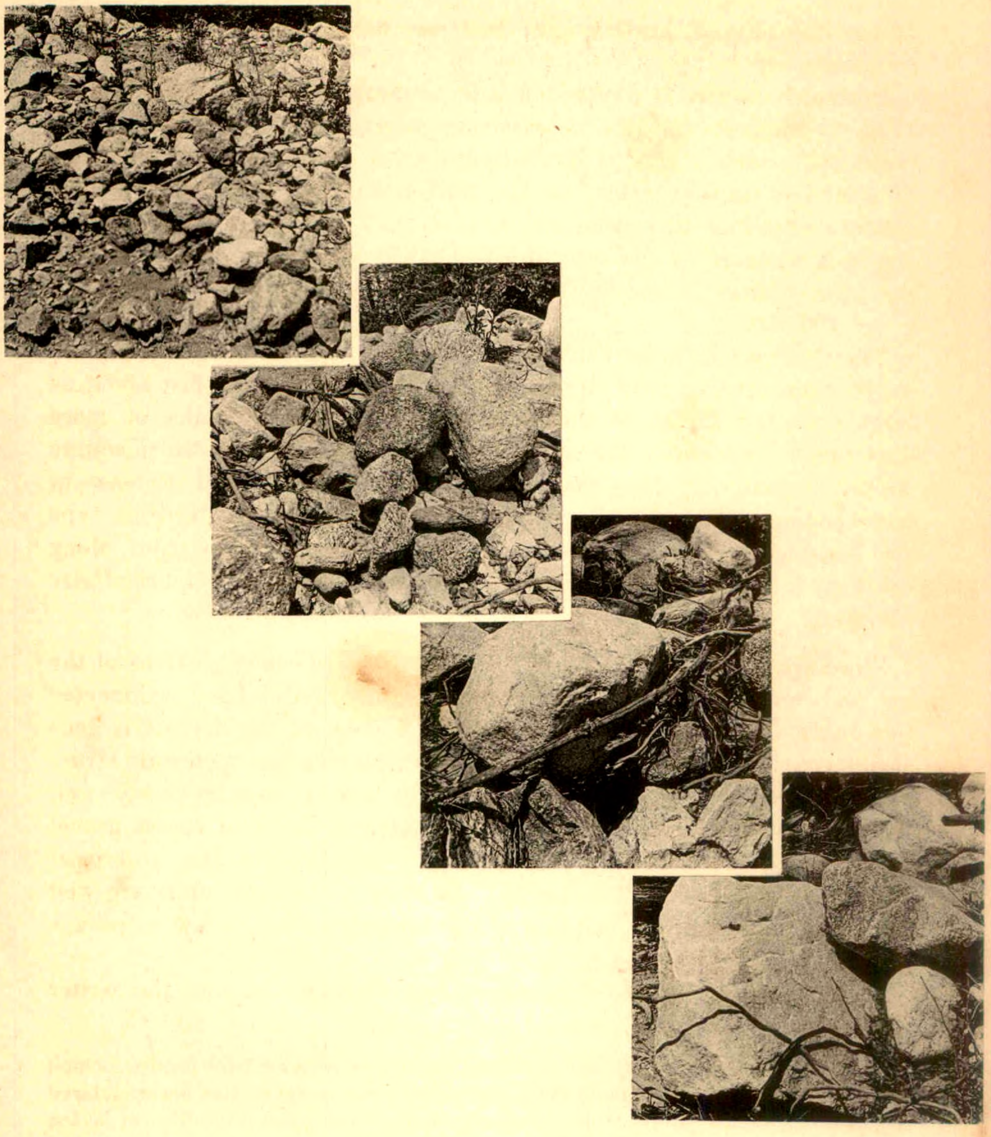
Inasmuch as size is plotted on a logarithmic scale, the straight lines (Fig. 7) suggest that the downstream increase in boulder size is exponential. Such a change downstream must require special conditions of selective transportation, as the size gradient is reversed with respect to what may be expected along a stream. The slope of the straight line is a measure of the rate of size change and in the lower parts of the jams at sites 21 and 25 there is a doubling of the mean boulder size every 100 feet.

The decrease in large boulder size was not regular from one jam to another downstream; the largest boulders in the jams were just about as large near the mouth of the canyon as they were 10 miles or more upstream. Apparently the numerous tributaries furnished an adequate supply of material along the entire canyon so that no net decrease in large boulder size took place. When boulders of a particular rock type are considered, however, the mean size decreases downstream along the canyon. Thus boulders of Lowe granodiorite show a marked size variation.

Structure of jam deposits.—In the central, well-bedded portions of the log and boulder jams are usually three rather well-defined sedimentation units (Otto, 1938). The lowermost portion of the deposit is generally coarse gravel with cobbles, often displaying an openwork structure. Above this is medium-sized sandy gravel, usually fairly well bedded and imbricated. At the top is a thinner layer of coarse gravel and cobbles,—apparently a lag deposit. Occasionally the lowermost unit is medium sand (Pl. 4, fig. 2). In Plate 4 the three units are well developed, although the bedding of the central portion is not so prominent as in some instances.

In describing the jam deposits in San Gabriel Canyon, the writer concluded (1940):

> "Some of the coarse lower material may consist of boulder bars formed behind fixed obstructions in the channel floor. In many cases, however, this lower material probably represents debris which dropped immediately upon initiation of a log jam. A considerable portion of finer material near the stream bed was entrapped with it, resulting in a poorly sorted deposit. This lower sedimentation unit may extend indefinitely upstream because of the slackwater effect produced by the temporary damming. Meanwhile, heavy deposition continued in the ponded areas near the log jam, the debris being supplied by the partially 'cleared' waters which continued to pour in from upstream. In this manner the intermediate well-bedded sedimentation unit accumulated. The upbuilding of the beds forced the stream to cut laterally around the jam unless the jam itself was partially undermined by the clear water pouring over its edge. In either case a channel lower than the surface



SIZE VARIATION ALONG JAM DEPOSIT

Photos taken about 35 feet apart along jam near site 21. (Downstream direction from top to bottom.) The largest boulder is about 4 feet in diameter. All photos to same scale.

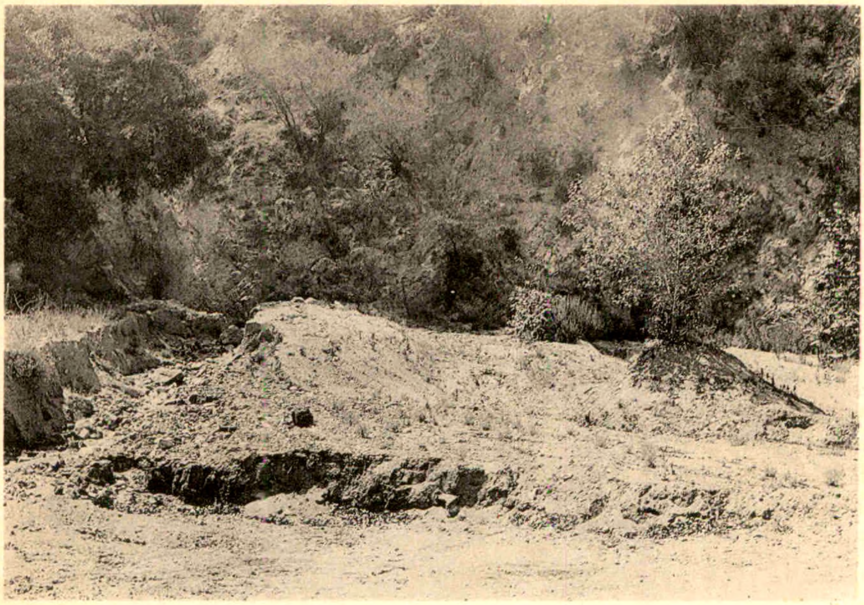


FIGURE 1. DEBRIS CONE FROM SMALL TRIBUTARY NEAR SITE 17

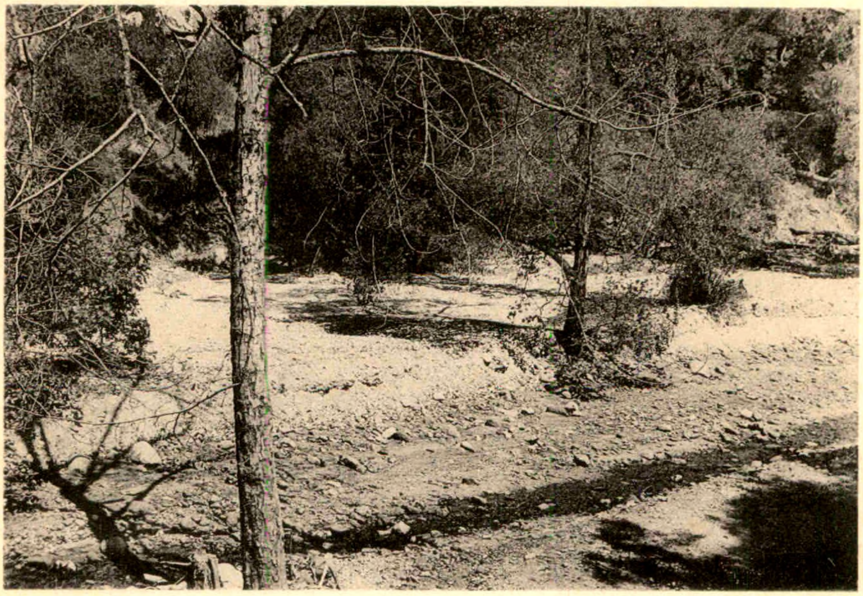


FIGURE 2. ALLUVIAL PLAIN DEPOSIT NEAR SITE 9
Post-flood erosion has developed a low terrace in these deposits.

NONJAM DEPOSITS IN ARROYO SECO

> of the deposit was soon developed. As the water collected into this new channel some of the topmost layers were swept downstream, leaving behind a lag deposit of cobbles and boulders, which is the topmost sedimentation unit."

This study supports the above conclusion, although some variations in the conditions of sedimentation may have occurred. The lower sedimentation unit was composed of poorly sorted gravel, of openwork seams, or of medium sand. The sand was found immediately upstream and downstream of a narrow gorge and may represent slackwater deposits formed during rising stages of the flood.

DEPOSITS AT TRIBUTARY MOUTHS

Deposition in the canyon was also associated with the mouths of many tributaries. Most of the tributaries have very steep gradients, and those entering from the west below Oakwilde contributed much debris, in part from the highway fills above. One prominent cone is at the mouth of Brown Canyon (site H, Fig. 2), another at the mouth of a minor tributary near site 17, leading directly down from the highway above (Pl. 6, fig. 1). Some tributaries supplied material from the highway fill, others carried large amounts of debris down from the burned-over area (Fig. 4). Fern and El Prieto Canyons, for example, had marked cones near their mouths composed of a rather heterogeneous mixture of fine and coarse material, with no evidence of sedimentation units such as those associated with the jams.

Immediately downstream from the tributaries the lithology of the main flood deposits usually reflected the contributions from the tributaries. In at least two instances the waters of the tributary and main stream evidently moved downstream side by side for an appreciable distance. The most marked occurrence of this sort was associated with Brown Canyon. This tributary contributed so much material to the main valley that the latter was literally choked; remnants of a large debris cone were plainly visible. This nonmixing of waters was discovered accidentally while pebble counts were being made. Along the main valley the pebble counts were expressed in terms of their content of Lowe granodiorite pebbles, which were used as an index of contamination. (The full discussion of pebble lithology is given later.) At site 11, 34 per cent was granodiorite pebbles, and at site 12 only 6 per cent. Additional samples from several local sites showed a marked and systematic variation in the content of granodiorite pebbles along this part of the valley. Figure 8 is a map of the field relations. Counts were made at the 29 sites indicated by numbers, which show the percentage of granodiorite pebbles at each point. The area studied was limited by the continuity of the deposits. At the downstream end the temporary road and shifting channels of the stream rendered it doubtful whether reliable counts could be had, and upstream, near the mouth of Brown Canyon, tractoring had destroyed the continuity of the deposits.

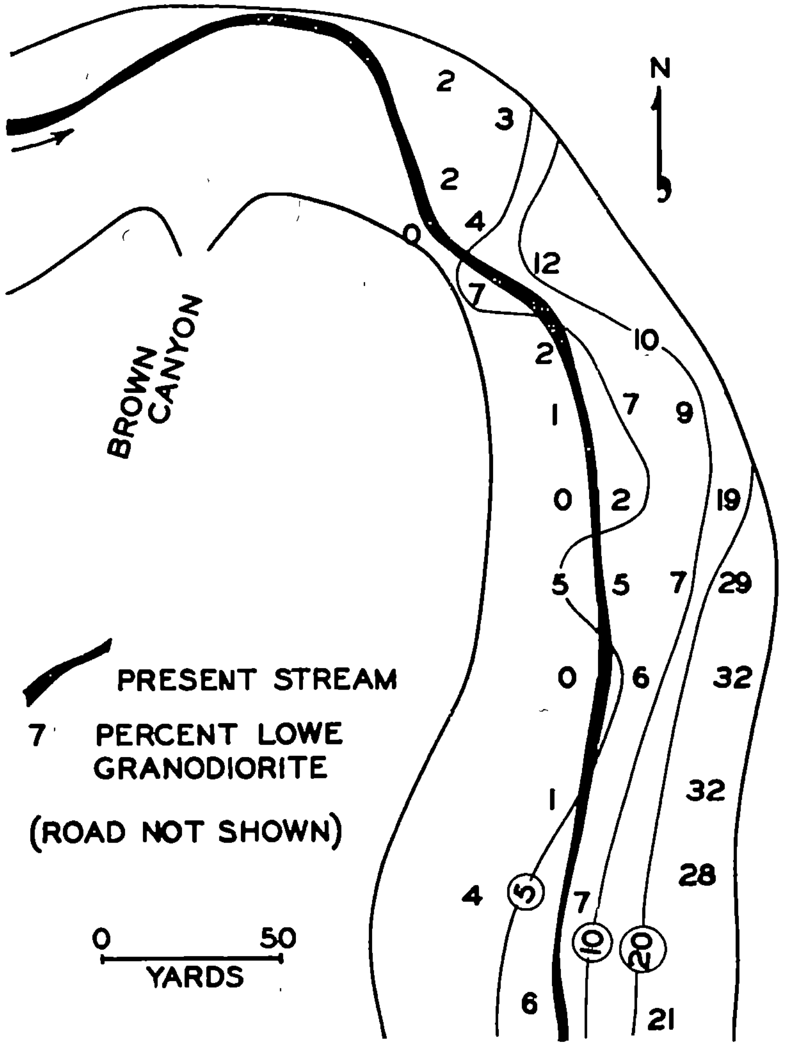


FIGURE 8—*Map of lithologic variation in deposits at site 12*

Showing effects of nonmixing of tributary and main stream waters. The figures refer to percentage of Lowe granodiorite pebbles at each sampling site.

The rapid change in granodiorite pebble content is reflected by the isolines, which are geometric in interval. The linear pattern in Figure

8 reflects the parallel movement of tributary and main stream waters below Brown Canyon. Along the west side granodiorite pebbles were rare or absent, whereas along the east side they were abundant. The mixing of the waters was apparently confined to a narrow central zone. The geometric spacing of the isolines indicates the relative abruptness of the change from none to about 30 per cent of granodiorite pebbles.

In the northern part of the map the granodiorite pebbles are almost entirely lacking over the entire width of the valley for a limited distance, in definitely undisturbed deposits. Possibly the large debris cone from Brown Canyon may have effectively choked the main valley, forcing the main flood waters, already loaded, to the north wall of the canyon. This constriction in the valley may have increased the velocity of the main flood waters as in a conduit, with the result that transportation occurred through the conduit without significant deposition. As the valley widened around the bend, the flood waters spread out, mixed in part with the heavily charged waters from Brown Canyon, and began deposition.

Zones of delayed mixing probably occurred at several tributaries, inasmuch as a number of them contributed significant amounts of debris, especially from Angeles Crest Highway. Below Dark Canyon, where delayed mixing also took place the deposits were too discontinuous for detailed mapping. As with Brown Canyon, the granodiorite pebbles increased markedly from west to east.

DEPOSITS IN BROAD PARTS OF CANYON

In the lower portion of Arroyo Seco, where the flood plain was present, the relatively broad and straight stretches were covered with a more or less uniform plain of sandy gravel. One of the best developed of these alluvial plains was found at site 9 (Fig. 2). Just upstream of the site is a sharp bend in a narrows, whereupon the canyon becomes relatively broad and straight for about a quarter mile. Figure 2 of Plate 6 shows the deposits, now present as a terrace caused by post-flood erosion. At the downstream end of the stretch is another sharp bend and narrows, but no marked log or boulder jam. The main deposit apparently thins somewhat above the narrows, descending along a steeper gradient to a zone of relative scour with lag boulders which occurred in the narrows associated with the bend.

The deposits in the straight portion showed three sedimentation units as in the jam deposits, so that very likely similar conditions of deposition prevailed, although the ponding may not have been so abrupt as in true jams. Instead, the downstream narrows controlled much of the deposition, for it increased the depth of water to accommodate the discharge, with the result that a backwater effect extended upstream, reducing the velocity and causing deposition. Meanwhile the rapidly moving partially cleared waters which entered the narrows produced a scour zone.

The three sedimentation units in this alluvial plain can apparently be related to early boulder bars or "pavement" over which the flood waters moved, and which now forms the bottom unit. The central bedded unit was formed during the bulk of sedimentation, and the topmost pebble and cobble lag was left behind by later waters which selectively reworked the top of the deposit.

The comparative simplicity of the picture at site 9 was elsewhere complicated by jam deposits which may have formed in conjunction with the alluvial plains in other parts of the canyon. Although the pronounced jams were often associated with bends, they sometimes occurred along a relatively straight stretch, where slight narrowing or other factors may have allowed a locking of logs or boulders. Therefore, the flood deposits presented a heterogeneous picture at first glance, which gradually resolved itself into the several elements as field work progressed. In some instances it was not possible to decide definitely which type of deposition predominated.

The canyon as a whole therefore presented a series of partially discontinuous deposits in its upper portion, succeeded by more continuous deposits downstream, with local regions of scour, and composed of partially distinct, partially overlapping deposits of jams, alluvial plains, and tributary-formed cones.

LITHOLOGY, TEXTURE, AND IMBRICATION OF FLOOD DEPOSITS

COLLECTION OF SAMPLES

To supplement the field data with more quantitative information, samples were collected for detailed laboratory study. Pebble counts were made at sites along the main valley and at every tributary mouth. Mechanical analyses were made along the lower portion of the canyon, where the deposits are well developed. Shape and roundness studies were made on pebbles downstream from Switzer's Camp (Fig. 2) to the canyon mouth. Fabric analyses were made of 11 samples, mainly in the lower part of the canyon.

Switzer's Camp was chosen as an arbitrary origin for purposes of graphing. It is in the central portion of the Lowe granodiorite area and is the highest point along the valley at which detailed analyses were made, with the exception of one oriented sample. All graphs given in later sections are referred to the origin at Switzer's Camp for convenience of comparison.

The samples were collected from the central, generally well bedded sedimentation unit at each sampling site. The samples for mechanical analysis were made by combining several samples through this unit at closely spaced intervals. Many of the samples were collected directly from jam deposits, but some of them were taken from cut banks in relatively straight parts of the valley, where the deposits were arranged in more regular alluvial plains. The following section summarizes the sampling data.

DESCRIPTION OF SAMPLES

Each item is keyed to indicate the type of sample taken. M = composite channel sample for mechanical analysis; L = sample for pebble count; S = shape analysis; R = roundness analysis; and O = sample of 100 oriented pebbles for fabric analysis.

Sample 1. Cut bank near center of valley, 2.75 miles upstream from Switzer's Camp. L, O.

Sample 2. Small bank along stream at Switzer's Camp. L, S, R.

Sample 2a. Talus sample (16-32 mm.) from steep ravine in Lowe granodiorite at Switzer's Camp. S, R.

Sample 3. Cut bank 0.95 mile downstream from Switzer's Camp. L, S, R.

Sample 4. Bank along stream, 0 3 mile downstream from sample 3. L, R.

Sample 5. Cut bank 0.45 mile downstream from sample 4 L, R, S.

Sample 6. Bank 0 8 mile downstream from sample 5. L, R, S.

Sample 7. Cut bank 1.0 mile downstream from sample 6. L, R, S.

Sample 8. Bank near east wall, 1.0 mile downstream from sample 7. L, S, R.

Sample 9. Bank in low terrace near west wall, 1.0 mile downstream from sample 8. M, L, R, S, O.

Sample 10. Cut bank in low terrace just below Oakwilde, 0.3 mile downstream from sample 9. M, L.

Sample 11. Cut bank in jam deposit, 0 3 mile downstream from sample 10. M, L, O.

Sample 12. Bank in low terrace, 0.25 mile downstream from sample 11 M, L, R, S

Sample 13. Cut bank near east wall of valley, 0 25 mile downstream from sample 12. M, L, O.

Sample 14 Poor bank near center of valley flat, 0.25 mile downstream from sample 13 M, L, R.

Sample 15. Cut bank in jam deposit, 0.25 mile downstream from sample 14. M, L, O.

Sample 16. Bank in composite (?) jam, 0.2 mile downstream from sample 15. M, L, R, S.

Sample 17. Low bank in flood terrace, 0 25 mile downstream from sample 16. M, L, O

Sample 18. Cut bank in jam near center of valley, 0.25 mile downstream from sample 17. M, L, R.

Sample 19. Cut bank in jam, 0 25 mile downstream from sample 18 M, L. O.

Sample 20. Bank in jam deposit, 0.25 mile downstream from sample 19. M, L, R, S.

Sample 21. Bank in low flood terrace, 0.2 mile downstream from sample 20 M, L, O.

Sample 22. Poor bank near center of valley, 0.3 mile downstream from sample 21 M, L.

Sample 23. Cut bank near east wall, 0 25 mile downstream from sample 22. M, L, O

Sample 24. Bank in jam deposit, 0.2 mile downstream from sample 23. M, L, R, S.

TABLE 1.—*Size composition of Arroyo Seco flood deposits*

(Expressed as weight percentages)

Sample	Grade limits in mm.											
	128–64	64–32	32–16	16–8	8–4	4–2	2–1	1–1/2	1/2–1/4	1/4–1/8	1/8–1/16	-1/16
9	...	7 6	14 1	14 4	13 0	15 2	12 2	10 5	7.5	3 7	1 1	0.4
10	.. ..	5 5	10.0	13 4	14 5	13 5	13 6	11 5	9 5	5 7	2.0	0.8
11	. ..	14 2	19.5	17 0	9 4	9 3	8 2	8 3	7 5	4 2	1 7	0 7
12	. ..	9 2	12 8	11 9	9 5	15 5	16 1	12 5	6 4	3 2	1.6	1 3
13	....	12 0	18 0	10 1	7 8	11 0	12 5	12 6	9 1	4 3	1 7	0 9
14	...	9.1	23.2	14 1	9 1	8.3	8 0	11 4	10 4	4 5	1 4	0 5
15	.. .	7 0	11 9	12 8	11 6	13 4	15 0	13.6	8 2	3 9	1 6	1 0
16	..	7 2	19.0	10 7	8 3	10 8	13 3	14.7	8 7	4 2	1.9	1 2
17	.....	11.6	14.7	12 1	13 5	15 9	13 1	10 1	5 0	2 2	1 0	0 8
18	. .	8 0	17 7	16 4	10 6	12.6	10 9	9 0	7 0	4 8	2 1	0 9
19	. ..	12 4	20 9	14 6	9 0	10 3	11 4	11 0	5 7	2 5	1 2	1 0
20	.. .	9.3	15 0	13 9	9 3	10 0	12.6	12 7	9 7	5.3	1 5	0 7
21	3.6	9 5	11 9	12 2	9.4	10 2	14 1	15 0	8 2	3 9	1 4	0 6
22	...	17 5	17 0	15.6	11 1	9 0	5 5	6.3	8 5	5.9	2 4	1.2
23	.. .	21 7	18 8	12 0	7 6	8 6	8 2	8 3	6 6	4 7	2.3	1 2
24	3 2	10 4	20.0	15 4	11 2	7 7	8.0	9 1	7 6	5 0	1 7	0 7
25	.	12 1	15 9	16.7	12 5	9 0	8 7	9 9	8 7	4 7	1 5	0 3
26	...	10 1	14 7	11 9	8 1	9 9	11.1	14 5	10 7	5 9	2 2	0 9
27	. ..	11 6	11 9	13 7	10 8	12 8	13.3	12 1	8 0	3 5	1 4	0 9
28	1.0	10 6	14.5	15 2	12 4	14 0	10 9	8 3	5 5	4 3	2 4	0 9

Sample 25. Bank in jam deposit, 0.25 mile downstream from sample 24. M, L.

Sample 26. Cut bank in jam deposit, 0 25 mile downstream from sample 25. M, L, R, O.

Sample 27 Bank in jam deposit, 0 3 mile downstream from sample 26. M, L.

Sample 28. Cut bank near east wall, 0 25 mile downstream from sample 27. M, L, S, R, O.

Sample A Little Bear Canyon. L.

Sample B Bear Canyon L.

Sample C. Unnamed tributary, 0.1 mile downstream from sample 4. L.

Sample D. Unnamed tributary, 0.25 mile upstream from site 8. L.

Sample E. Unnamed tributary, 0.2 mile downstream from site 8. L.

Sample F. Dark Canyon. L.

Sample G. Twin Canyon L.

Sample H. Brown Canyon. L.

Sample J. Pine Canyon. L.

Sample K. Falls Canyon. L.

Sample L. Nino Canyon. L.

Sample M. Agua Canyon L

Sample N. Fern Canyon. L.
Sample P. Unnamed tributary, 0.1 mile upstream from site 26. L.
Sample Q. El Prieto Canyon. L.
Sample R. Millard Canyon. L.

TABLE 2.—*Statistical parameters of size data*

Sample	Phi mean M_ϕ	Phi standard deviation	Geometric mean diameter in mm. GM_ξ
9	−2.23	2.48	4.70
10	−1.41	2.37	2.66
11	−2.32	2.56	5.00
12	−1.70	2.42	3.25
13	−1.84	2.58	3.58
14	−2.04	2.57	4.11
15	−1.87	2.48	3.65
16	−1.63	2.54	3.10
17	−2.11	2.33	4.31
18	−1.93	2.46	3.81
19	−2.23	2.46	4.70
20	−1.69	2.54	3.22
21	−1.81	2.58	3.50
22	−2.27	2.74	4.81
23	−2.42	2.86	5.50
24	−2.38	2.66	5.21
25	−2.09	2.52	4.26
26	−1.63	2.66	3.10
27	−1.82	2.46	3.52
28	−2.05	2.50	4.14

MECHANICAL ANALYSIS

Twenty samples were collected for detailed mechanical analysis along the main stream from site 9 above Oakwilde to the canyon mouth, at site 28. The problem of choosing sampling sites was complicated by the apparent heterogeneity of the deposits and by the extreme size variations within the jam units. As in the earlier San Gabriel study, the sampling was confined to the central well-bedded portions of the flood deposits. The data thus obtained permitted a comparison of mean size and degree of sorting from jam to jam and along the interspersed alluvial plain deposits.

Each sample for size analysis weighed approximately 100 pounds. Material from the central sedimentation unit was collected from several closely spaced vertical channels through the unit to obtain a representative sample. The gravel, including all the coarse and fine material, was shoveled or scooped into canvas bags. These samples were analyzed into Wentworth grades.

The mechanical analysis data are given in Table 1. The particle size ranged from 128 mm. down to approximately $\frac{1}{32}$ mm. The amount of subsieve material (less than $\frac{1}{16}$ mm.) was generally so small that the range was not extended downward. Most samples have their maximum grade near the upper limit, and secondary maxima are not uncommon. Two statistical parameters were computed for each sample—the phi mean and the phi standard deviation (Krumbein and Pettijohn, 1938, p. 239-250). The phi mean was converted to its antilog, the geometric mean diameter in mm. These parameters are listed in Table 2.

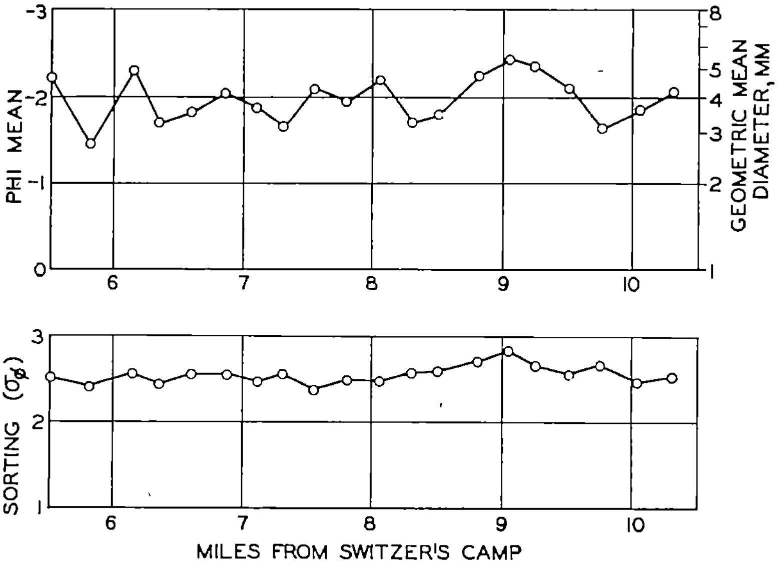


FIGURE 9.—*Variation in mean size and degree of sorting of flood deposits*

In central portions of jams and other flood deposits. The distance scale is referred to an upstream origin, for comparison with later graphs

Figure 9 is a graph of the variation of size and sorting along the valley. The broken line passing through the points does not strictly indicate the interpolated size between sampling sites, for each jam constitutes a unit, with considerable size variation from one end to the other. Interspersed among the jam units were more regular flood plain deposits in the straight and relatively broad parts of the valley. The samples were spaced approximately equally, so that they represent both kinds of deposits, depending upon which was present at a given sampling site. The present graph compares the central parts of succeeding deposits and as such shows a remarkable uniformity both in average size and degree of sorting from one deposit to the next. The absence of a marked trend indicates that variations are nonsystematic and that the individual flood deposits represent essentially similar dynamic responses to transporting conditions all along the canyon. These data bear out a similar conclusion reached in San Gabriel Canyon.

LITHOLOGICAL ANALYSIS

In the size samples pebble counts averaged about 250 pebbles each. For convenience of comparison the counts were confined to the 16-32 mm. range. The pebbles were first classified into half a dozen categories according to rock type and were then grouped into classes which reflect to some extent the known areal geology of the region. The final classification adopted is as follows:

DOMINANTLY LIGHT-COLORED ROCKS:

(1) Pink granite, coarse to fine texture, locally pegmatitic, porphyritic, or aplitic. Dark mineral usually biotite or hornblende, irregularly scattered through rock. Quartz prominent. This group includes the Echo granite of Miller's report (1934).

(2) White granodiorite, moderately coarse textured, locally gneissic Hornblende occurs in slightly elongated and somewhat irregular patches; in general, the dark minerals constitute less than 5 per cent of the rock. Small garnets fairly common in some facies. Quartz locally prominent. This is the Lowe granodiorite of Miller's report (1934).

(3) Gray granodiorite, coarse to fine texture, varying from quartz monzonites to quartz diorite. Dark mineral mainly hornblende; some biotite This group of rocks is distinctly different from the white granodiorite in appearance, and little difficulty is encountered distinguishing them in the field.

(4) Fine-grained igneous rocks, typically felsites.

DOMINANTLY DARK-COLORED ROCKS:

(5) Dioritic and other coarse-grained basic rocks, medium to coarse texture, dark minerals predominant.

(6) Fine-grained dark rocks, including basalt, dike rocks, and greenstone.

(7) Foliated rocks, mainly schistose, with mica schist predominant. Includes well-developed gneisses and other banded rocks, usually dark.

Of these several categories the most outstanding group, and one of the most easily recognized, is the white Lowe granodiorite. The percentage of these pebbles decreased gradually downstream from site 9, and, since this decrease apparently indicated successive dilution of an original load, the pebble counts were extended to include the entire stretch of the valley from the Lowe granodiorite area (Fig. 3) to the canyon mouth. These samples were supplemented by pebble counts made at every tributary mouth along the same stretch, to round out the picture of lithological variation. A total of 45 pebble counts were made, and the data are listed in Table 3 on a percentage basis. The counts at the tributaries were made far enough upstream from their mouths so that possible backwater effects were eliminated. The data from the tributaries (samples A to R) show that, beyond the area of Lowe granodiorite proper, no tributaries penetrated into the gran-

TABLE 3.—*Lithological composition of 16–32 mm. pebbles, Arroyo Seco*
(Expressed as number percentages)

Sample	Light-colored rocks			Dark-colored rocks			
	Pink granite	White (Lowe) granodiorite	Gray granodiorite	Felsite	Diorite	Basalt, etc.	Foliated rocks
1	12	1	32	0	10	2	43
2	10	5	59	1	9	3	13
3	16	19	45	2	4	4	10
4	11	32	38	1	6	5	7
5	9	34	33	3	6	4	11
6	10	45	28	0	4	3	10
7	9	42	32	2	2	1	12
8	2	41	18	3	11	6	19
9	8	34	17	6	17	7	11
10	14	33	25	5	12	3	8
11	11	34	20	6	16	3	10
12	7	27	28	4	16	4	14
13	16	31	25	4	15	5	4
14	18	26	30	3	14	5	4
15	17	20	39	2	15	4	3
16	10	22	28	4	18	9	9
17	8	27	31	7	14	5	8
18	21	21	24	4	17	5	8
19	20	19	27	4	20	4	6
20	18	21	27	3	20	5	6
21	18	22	26	3	18	8	5
22	15	24	30	3	14	6	8
23	28	19	28	1	13	4	7
24	18	23	31	3	13	4	8
25	18	23	29	2	15	4	9
26	20	21	27	6	12	7	7
27	18	23	28	3	11	7	10
28	26	20	23	3	13	5	9
A	1	82	13	0	1	2	1
B	20	40	35	2	0	3	0
C	9	59	5	0	4	4	19
D	2	30	27	3	27	5	6
E	14	0	50	1	10	2	23
F	25	0	53	0	16	2	4
G	27	0	37	0	28	6	2
H	22	0	55	0	14	5	4
J	17	0	39	0	16	3	24
K	27	0	50	tr	20	3	tr
L	12	0	80	1	5	2	0
M	15	0	77	0	5	2	1
N	15	0	41	2	7	5	30
P	12	1	66	11	3	6	1
Q	10	0	47	0	6	4	33
R	30	17	31	2	8	1	11

odiorite except the unnamed tributary at site D and Millard Canyon at site R.

The data of Table 3 show that the quantity of Lowe granodiorite pebbles (third column) increases successively in the upstream samples, reaches a maximum part way along the valley, and then decreases (Fig. 10). The rising part of the curve lies within the area of the Lowe granodiorite itself, and the maximum is reached at sample 6, near the western contact of the granodiorite. Within the Lowe granodiorite area practically all additions to the stream load are of that material, so that in passing through this section the load was continuously enriched by granodiorite pebbles. Beyond the downstream contact (with the exception of the tributary at site D) no more granodiorite was furnished by any tributary. The percentage of this rock thus diminished continuously as a function of the foreign contributions along the rest of the valley. The curve of decreasing granodiorite-content levels off downstream to an asymptote of about 20 per cent. This may indicate an approximate equilibrium between contributions from tributaries and the load coming down the main valley. Because the data are confined to the 16-32 mm. size range, however, this relation may be modified by a flow of material from one grade to another due to breakage and wear en route.

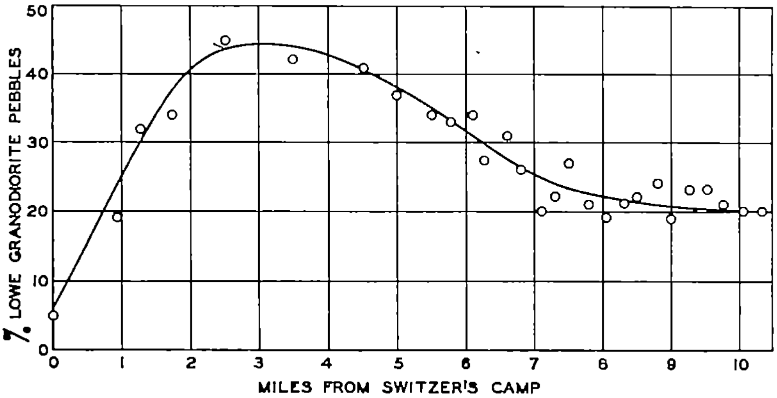


FIGURE 10—*Variation in percentage of Lowe granodiorite pebbles in flood deposits along Arroyo Seco*

SAMPLING ERRORS OF SIZE AND LITHOLOGY ANALYSES

The samples for size analysis in the San Gabriel study averaged about 15 pounds each, and the question was raised whether these adequately represented the coarser sizes. In attacking this problem a series of samples was collected both in San Gabriel Canyon and Arroyo Seco, ranging from about 15 pounds to 300 pounds. Invariably the larger samples contained larger pebbles and cobbles, but in proportion to the total weight of the sample these had relatively little effect on the size parameters. In a 300-pound sample from Arroyo Seco, for instance, was a cobble weighing 8.5 pounds, and yet, when the entire analysis was taken into account, the effect of this cobble on the median diameter was very slight. A 50-pound sample taken from the same site differed in its median diameter by less than 5 per cent. A comparative study of a set of six closely spaced samples in San Gabriel Canyon, of approximately the same weight as the first year's samples, showed a probable error of sampling of approximately 11 per cent. This is smaller than the observed variation from sample to sample along the canyon, but to reduce the sampling error it was concluded that samples twice as large would prove more satisfactory. The present size study in Arroyo Seco was based on 100-pound samples, for which the probable error of sampling is estimated to be less than 5 per cent.

TABLE 4—*Size variation of Lowe granodiorite boulders*

(Numbers represent arithmetic mean of 5 largest)

Distance downstream from Switzer's camp	Average size, meters
1.00	1.49
1.25	1.83
1.75	1.34
2.25	1.28
2.50	1.43
3.00	1.34
3.25	1.13
3.75	1.03
4.25	1.03
4.50	0.98
4.75	.76
5.00	.89
5.40	.79
6.00	.67
6.35	.61
7.10	.67
7.50	.55
7.90	.58
8.50	.52
8.70	.46
9.10	.55
9.60	.49
10.20	.58

The probable error of pebble counts was studied in a set of 24 closely spaced samples at site 9. The total probable error was found to be 10.2 per cent for sets of 100 pebbles, in terms of their Lowe granodiorite content. Assuming this same order of variability along the canyon, the samples of from 200 to 300 pebbles which were used for the data of Table 3 reduced this error to about 6 per cent.

SIZE VARIATION OF LOWE GRANODIORITE BOULDERS ALONG THE CANYON

A special supplementary study was made of the Lowe granodiorite boulders themselves. Boulders of all compositions in the jams are about the same size all along the canyon. It was anticipated, however, that if measurements were made of Lowe granodiorite boulders themselves there should be a marked decrease in size down the valley, due in part at least to continued breakage and abrasion.

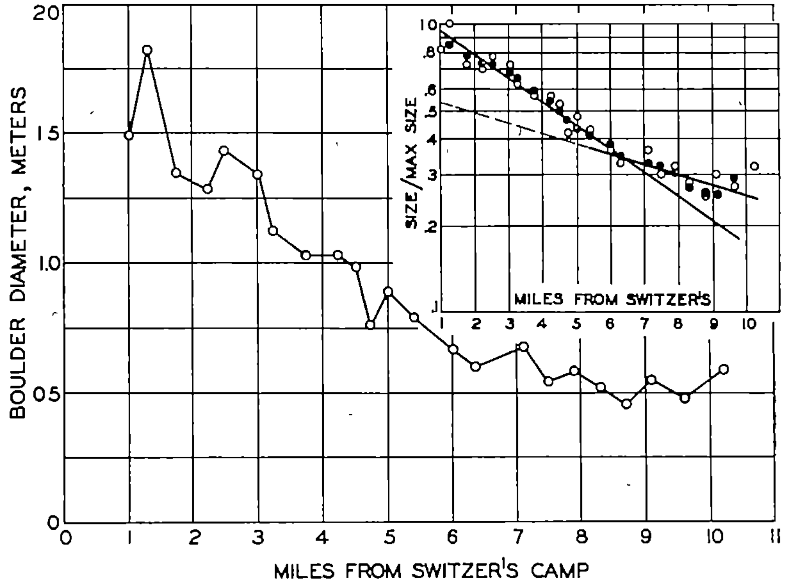


FIGURE 11.—*Variation in size of Lowe granodiorite boulders along Arroyo Seco*

Inset shows same data on semi-logarithmic plot. The black dots are moving averages based on successive sets of three observations

To obtain pertinent information Lowe granodiorite boulders were measured at intervals along the canyon. In the upper reaches the boulders were very numerous, but downstream they became comparatively scarce. For comparison, therefore, a number of boulders were measured at each site, and the average of the five largest was taken as an index. Table 4 lists the data, and Figure 11 shows the results. The data scatter rather widely, but a definite trend is discernible. The exponential trend of the data is brought out by the inset figure, which shows boulder size plotted on a logarithmic scale. The solid circles represent a moving average based on successive sets of three observations. Two straight lines are drawn through the averaged data; these suggest a decrease in the rate of change of boulder size about 6 miles downstream from Switzer's Camp (and about 1 mile downstream of the point where the flood plain appears in Arroyo Seco). If the break in the curve does coincide approximately with the transition region over which, on the average, scour conditions are replaced by depositional conditions, the change of slope in the graph may have a physical significance. Scour conditions imply more vigorous wear and breakage, whereas depositional conditions imply relatively more selective transportation. Thus the difference in slope of the two lines may offer a first approximation to the influence of the breakage factor during transport.

TABLE 5—*Sphericity data of Lowe granodiorite pebbles, 16–32 mm. class*

(Frequency expressed as number percentages)

Sample	.40–.49	50–.59	.60–.69	.70–.79	.80–.89	90+	Mean sphericity	Standard deviation
2	2.2	11 1	53 3	26 7	4 5	2 2	0 68	0.088
3	2 6	13 1	29 0	47.4	7 9	. .	.70	.091
5	2 9	14 3	40.0	34 3	7 1	1 4	.68	.095
6	1 6	23 8	42 9	25 4	6 3	. ..	68	086
7	.....	12 2	34 7	40 9	12 2	. ...	.70	.086
8	1.3	6.4	38 5	42.3	11.5	.. .	.71	.082
9	.....	10 0	36 7	41 1	11.1	1.1	.71	.086
12	....	14.0	28 1	42 1	15.8	......	71	092
14	2.4	3.6	36.1	39 8	18 1	.....	.72	089
16	2.4	16 6	34 5	40 5	4.8	1 2	68	.093
20	....	5 8	37.9	46 0	9 2	1.1	71	079
24	2.2	12 2	41 2	32 2	12 2	. ...	69	.099
28	... .	17.8	32.2	43 6	4 8	1.6	.69	.092

Strictly speaking, this analysis of boulder size goes beyond the effects of a single flood; probably any given boulder moves only a relatively short distance during a single flood, so that the observed size variation is the cumulative effect of a number of movements. The size variation is a composite of the effects of selection during transport, breakage, and abrasion en route, and perhaps of scour by finer material while the boulders are at rest (Maxson, 1940). That one of these factors predominates in given parts of the canyon is likely, and with adequate data it should be possible to isolate some of the variables.

SHAPE AND ROUNDNESS ANALYSIS

San Gabriel Canyon carried a load which lithologically had much greater uniformity than did Arroyo Seco where numerous tributaries contributed a large part of the load. Observation as well as theory indicate large variations in shape and roundness along Arroyo Seco, even within a given size range. Nevertheless, the Lowe granodiorite pebbles offered a means for studying both shape and roundness as functions of distance along the stream, because any pebbles in the lower part of the canyon could have reached that point only by traveling along the main valley. The situation is thus similar to Wentworth's Russell Fork study (1922).

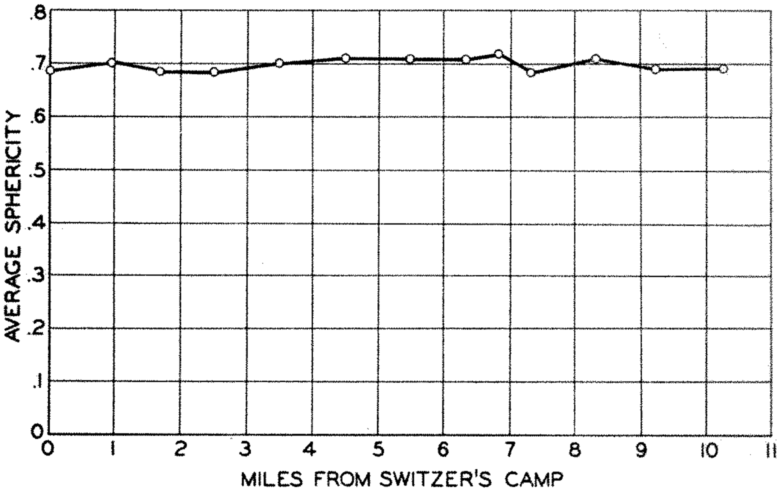


FIGURE 12.—*Variation in pebble shape (sphericity) along Arroyo Seco*
Lowe granodiorite, 16-32 mm. class.

Some of the samples collected for mechanical analysis contained sufficient pebbles for shape and roundness study. These were supplemented by samples collected upstream as far as Switzer's Camp to bring out the essential features of the curves. The shape (sphericity) was measured by a graphic intercept method based on axial ratios of the pebbles. This method is more rapid than earlier methods and gives results in substantial agreement. Roundness was estimated by a rapid visual method from a standard set of pebble images. Both methods are described in detail elsewhere (Krumbein, 1941a).

The sphericity data are listed in Table 5, which also includes the arithmetic mean sphericity of the samples. Figure 12 shows the variation in mean shape along the valley. The remarkably uniform values from point to point indicate a small nonsystematic variation, reflecting uniform conditions of shape selection along the valley, and thus indicating essentially similar dynamic conditions of deposition. The same feature was shown, but less clearly, in San Gabriel Canyon.

Figure 13 also brings out the absence of any marked change in pebble sphericity over the distance studied. The ratio b/a is the numerical value obtained by dividing the length of the intermediate pebble axis by the longest pebble axis. The ratio c/b is the ratio of the short axis to the intermediate axis. The curves represent lines of equal sphericity superimposed on the axial ratios (Krumbein, 1941a). The dashed lines are Zingg's shape classes (Zingg, 1935; Krumbein and Pettijohn, 1938, p. 288). Comparison of the two figures shows that there is no marked change in sphericity between the two sets. In both instances the majority of the pebbles have sphericity values between 0.6 and 0.8. The centers of gravity of both samples (shown by the triangles) lie at almost exactly the same positions. The degree of scatter of the data is also roughly the same.

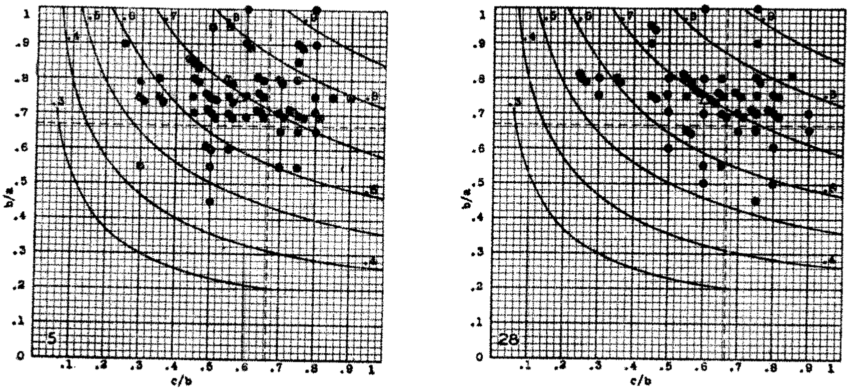


FIGURE 13.—*Comparison of sphericity distributions of samples 5 and 28*

See text for details of plotting.

The roundness data offered several difficulties, especially in the upper part of the canyon. Within the area of Lowe granodiorite new material was always available to the stream, so that throughout that stretch angular fragments were continuously added to a load of already partially rounded granodiorite pebbles. Once the stream leaves the granodiorite area no more granodiorite fragments are contributed,[1] so that the average roundness tends to become more uniform. The result is an inevitable scatter of the data among the upstream samples. To reduce this scatter, an average was taken of samples 2, 2a, and 3, which represent the first three samples in the granodiorite area. Sample 2a came

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[1] An exception is the unnamed tributary at site D, but even here the pebbles have traveled a fair distance from the outcrops before they enter the main valley.

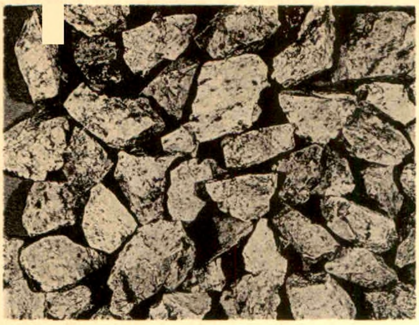

FIGURE 1. FRAGMENTS FROM TALUS AT SWITZER'S CAMP

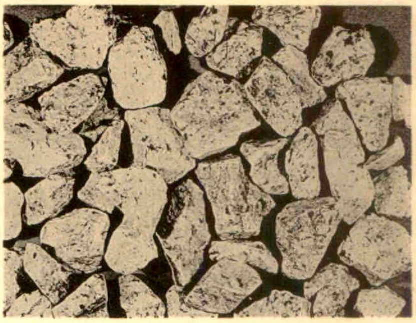

FIGURE 2. PEBBLES FROM CUT BANK AT SWITZER'S CAMP

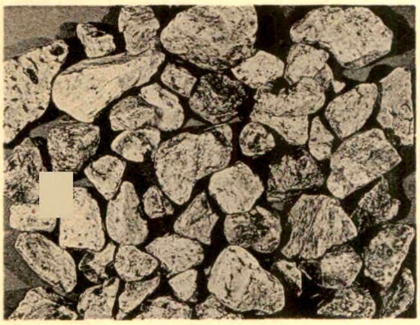

FIGURE 3. SAMPLE 7, 3.5 MILES DOWNSTREAM FROM SWITZER'S CAMP

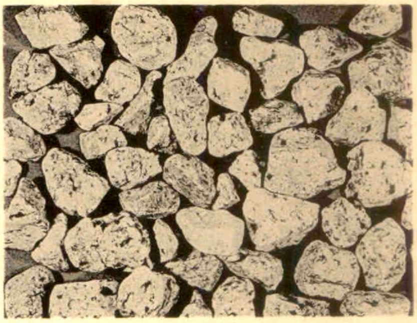

FIGURE 4. SAMPLE 8, 4.5 MILES DOWNSTREAM FROM SWITZER'S CAMP

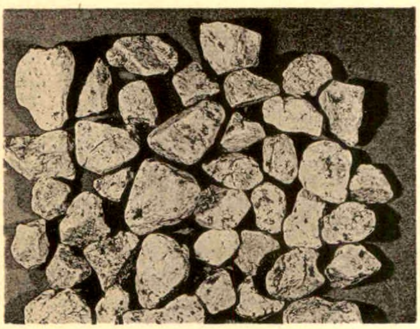

FIGURE 5. SAMPLE 22, 8.8 MILES DOWNSTREAM FROM SWITZER'S CAMP

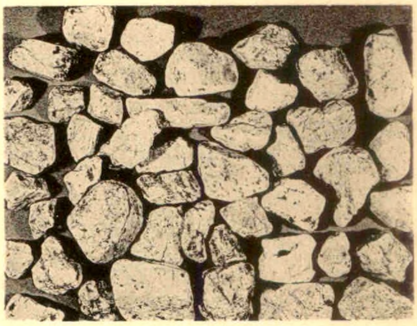

FIGURE 6. SAMPLE 27, 10.0 MILES DOWNSTREAM FROM SWITZER'S CAMP

ROUNDNESS VARIATION IN LOWE GRANODIORITE PEBBLES, 16–32 MM. CLASS from a talus slope along the canyon and gives an indication of the initial roundness of the fragments.

Plate 7 shows the roundness variation in six samples. The total change is not very marked, but differences among the first few samples are greater than among the later ones, indicating more rapid rounding during the early stages of the process.

TABLE 6.—*Roundness data of Lowe granodiorite pebbles, 16–32 mm. class*

(Frequency expressed as number percentages)

Sample	Midpoints of roundness classes						Mean roundness
	.1	.2	.3	.4	.5	.6	
2	45.2	22.7	22.7	9.4	....	....	0.20
2a	98.5	1.5	....	....	....	....	.10
3	7.9	63.1	29.0	....	....	....	.23
Composite*	57.5	24.2	15.0	3.3	....	....	.16
4	35.8	43.4	18.9	1.9	....	....	.19
5	17.6	50.0	23.0	8.1	1.3	....	.23
6	5.8	55.0	29.0	7.3	2.9	....	.25
7	....	28.3	41.5	26.4	3.8	....	.31
8	....	8.3	50.4	36.3	5.0	....	.34
9	....	6.0	56.0	36.0	2.0	....	.34
12	....	2.2	51.1	44.5	2.2	....	.35
14	....	12.2	48.2	30.9	6.2	2.5	.34
16	....	....	57.1	37.2	5.7	....	.35
18	....	7.0	27.9	57.0	8.1	....	.36
20	....	2.6	41.0	48.6	7.8	....	.36
24	....	....	57.0	41.4	1.7	....	.35
26	....	....	42.1	50.0	7.9	....	.37
28	....	2.8	28.7	51.4	17.1	....	.38

* Average of samples 2, 2a, 3.

The roundness data are listed in Table 6, and the corresponding graph of arithmetic mean roundness is shown in Table 14. The data show a definite trend, leveling off toward an asymptote of about 0.40 downstream. To bring out this asymptotic value more clearly, several additional samples were measured near the lower end of the series. The inset of Figure 14 shows the roundness data plotted on semilogarithmic paper, assuming an asymptote of 0.40. The ordinates represent the difference between the asymptote P_o and the value of the roundness P, divided by the asymptotic value. The straight line indicates the fair agreement of the data with an exponential trend, as discussed in the San Gabriel paper (Krumbein, 1940). The coefficient of rounding in the present instance is 0.23 as compared to 0.50 in San Gabriel.

As in the San Gabriel deposits, the roundness of the pebbles is a function of the distance transported. The smaller coefficient of rounding may be the result of one or more factors. The rocks involved may have had different resistances to wear; the velocity of the flood waters may have been different; and the Arroyo Seco pebbles were from a slightly smaller size range than the San Gabriel pebbles. The absolute magnitude of the coefficients is perhaps not so significant as their relative values. The ratio 0.50/0.23 = 2.2 means that on the average the San Gabriel pebbles were rounded twice as rapidly as those in Arroyo Seco.

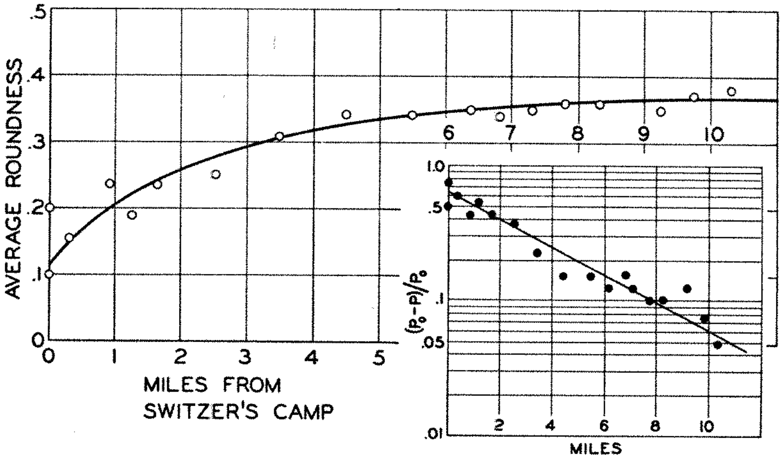

FIGURE 14.—*Variation in pebble roundness along Arroyo Seco*

Lowe granodiorite, 16-32 mm. class. Inset shows same data on semi-logarithmic plot.

Neither the San Gabriel nor the Arroyo Seco data show the incipient steeper portion of the roundness curve found experimentally (Krumbein, 1941b). Field conditions may obscure this feature, although it was present in Wentworth's Russell Fork data, discussed in the abovementioned paper. Possibly, in the distances covered in Arroyo Seco and San Gabriel Canyon, the accelerated rate of rounding during the flood may extend the initial steep portion over a greater distance than under less rigorous conditions.

IMBRICATION OF PEBBLES IN FLOOD DEPOSITS

During the San Gabriel study, evidence was found of reverse imbrication in the sample collected at the highest upstream position. The general site was revisited during 1940, and several additional oriented samples were collected. Upstream from the original site a very large boulder set up a "shadow zone" of reversed imbrication on its down-

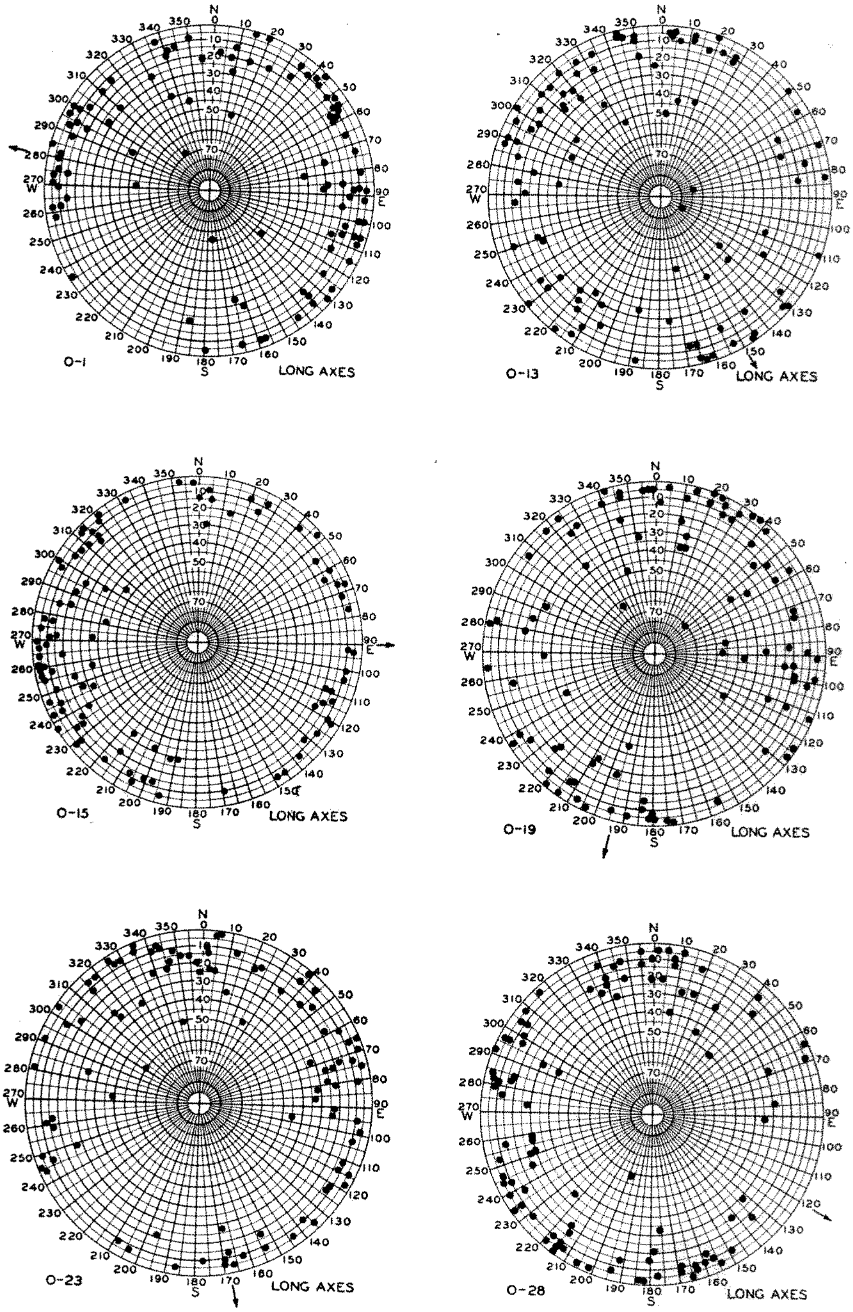

FIGURE 15.—*Polar co-ordinate charts of pebble orientation in Arroyo Seco*

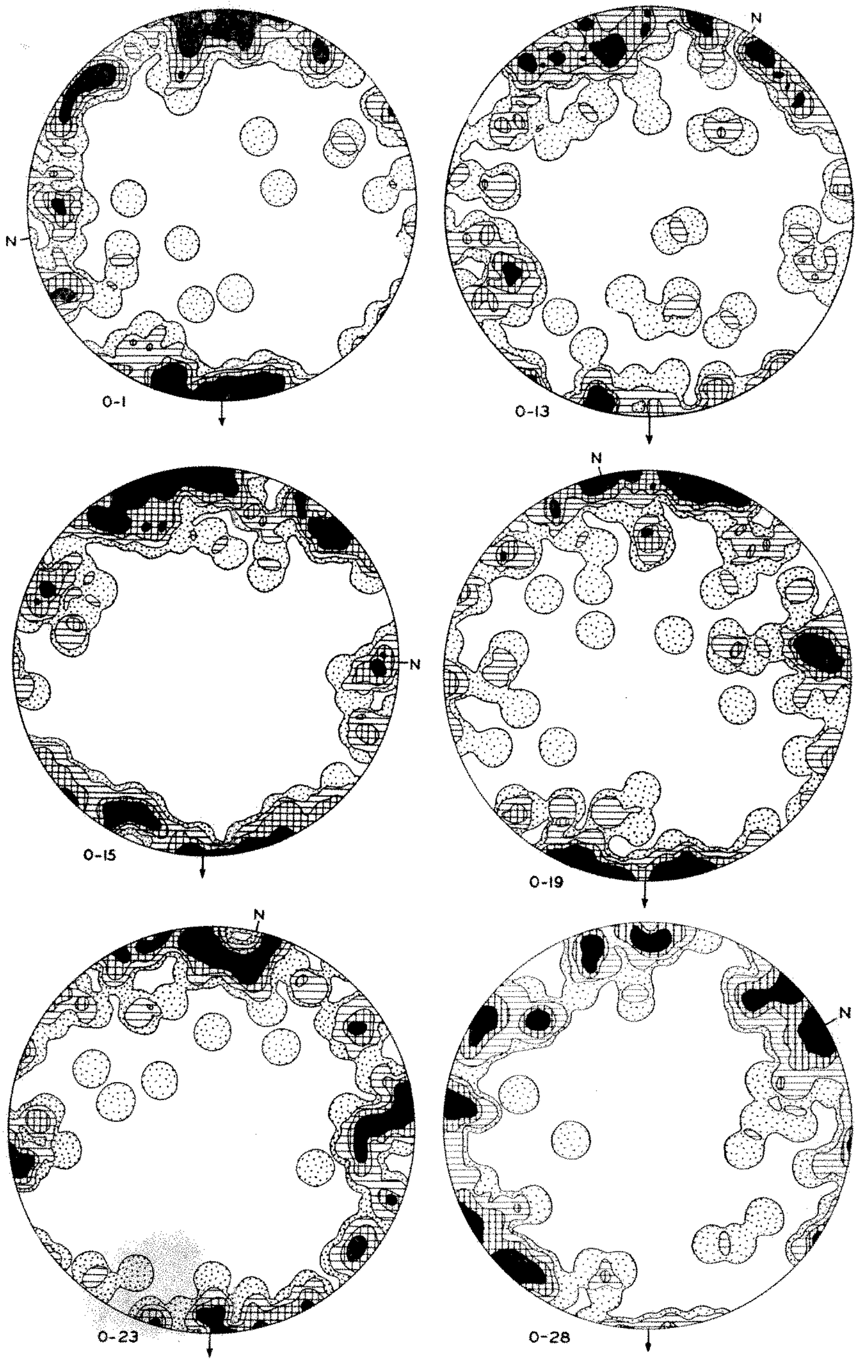

FIGURE 16.—*Petrofabric diagrams of Arroyo Seco pebbles*

stream side, extending nearly 150 feet and tapering off. Apparently sufficient disturbances were set up in the lee of the boulder to generate reverse currents or eddies, in the zone of which the deposited materials were imbricated by a reverse current. Other similar large boulders in the canyon indicated that the magnitude of the "shadow zone" was a function of the boulder size.

TABLE 7.—*Dip direction and average roundness of oriented pebbles (Arroyo Seco)*

Sample	Mean azimuth	Azimuthal standard deviation	Mean roundness
0- 1	106 2°	43 8°	0 28
0- 9	3 6°	46 4°	32
0-11	51 0°	46 4°	31
0-13	143.0°	46.0°	23
0-15	79 2°	43 2°	19
0-17	136 6°	46 2°	24
0-19	21 0°	47 4°	.25
0-21	119 8°	51 0°	32
0-23	155 4°	49 4°	34
0-26	179 2°	48 2°	32
0-28	168 2°	47 8°	32

In Arroyo Seco, by means of an orienting frame (Krumbein, 1939), one oriented pebble sample was collected in the steep upper part of the canyon (site 1, Fig. 2), and 10 samples were collected in the lower part of the valley, from sites 9 to 28. In every instance a normal upstream imbrication was found; Figure 15 shows the data of six typical samples. In Figure 15 any given point has the co-ordinates, dip direction (measured along a circumference), and angle of dip (measured inward along a radius). Figure 16 shows the corresponding petrofabric diagrams, all oriented so that the direction of valley trend is vertically downward. The diagrams show normal imbrication, including sample 0-1, which was collected at a point where the stream gradient was about 500 feet to the mile. In each instance the symmetry is monoclinic, with the *a*-fabric axis more or less parallel to the valley trend. Considering the numerous swirls, eddies, and reverse currents that may be present in floods, the general agreement of the preferred orientation to valley trend is significant. Apparently the mean azimuth may be safely considered as a statistical index of direction of stream flow.

In the San Gabriel study the azimuthal standard deviation of the pebbles about the mean preferred orientation varied as a function of distance along the valley, but the present data do not show any systematic variation downstream. It was also noted in the earlier study that the azimuthal standard deviation was larger for rounded pebbles than for angular pebbles. The Arroyo Seco data do not clearly support this observation. Table 7 lists the azimuthal standard deviation of the 11 Arroyo Seco samples, as well as the average roundness of the oriented pebbles. The pebbles in these samples include all rock types, so that the average roundness does not agree in general with the average roundness of the Lowe granodiorite pebbles given in Table 6. Figure

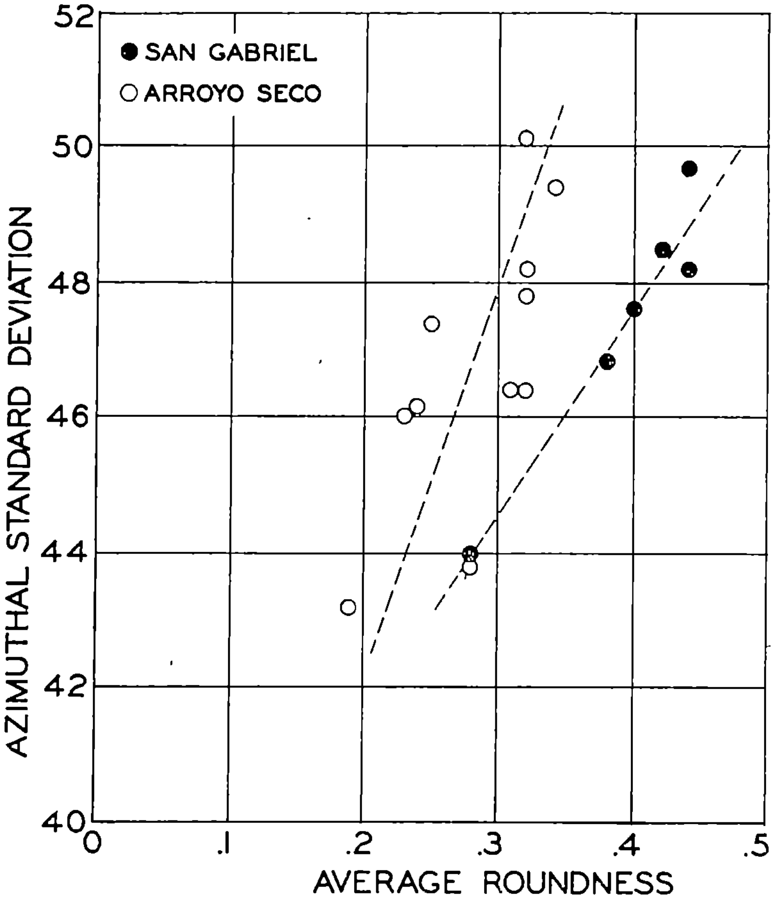

FIGURE 17—*Scatter diagram of average roundness and azimuthal standard deviation of oriented pebbles*

Arroyo Seco and San Gabriel Canyon.

17 is a graph of azimuthal standard deviation plotted against mean roundness. The open circles represent the Arroyo Seco samples, and the solid circles represent the San Gabriel samples (Krumbein, 1940, Tables 6, 7). The Arroyo Seco data scatter rather widely, whereas the San Gabriel data show a fair linear trend. In view of this unsatisfactory evidence, generalizations regarding the spread of pebble axes as functions of distance or of average roundness are deferred.

INTERPRETATION OF FIELD AND LABORATORY DATA

GENERAL CONSIDERATIONS

Some interpretations have been given in the descriptive portion of the paper, but an integration of the several lines of evidence should be made. Three general problems deserve discussion: (1) the mechanics of transportation of flood materials; (2) the physical factors which control their deposition; and (3) the interplay of wear and selective transportation during movement. These features must be discussed almost entirely from post-flood evidences as direct observational data on floods in steep canyons are very meager. Engineers are naturally concerned most with damage to business and residential districts, which confines their work during floods to regions beyond the canyon mouths. Stream gauges and other equipment along the canyons are usually destroyed or ineffective during large floods, so that many basic data are never obtained. Estimates of discharge based on high water marks are at best uncertain.

The geological approach offers the best solution until more quantitative observational data are available. Gradational processes are reflected in the characteristics of the deposits they form, and hence a study of the deposits should shed light on the mechanical processes involved.

Troxell and Peterson (1937), studied the effects of bends and constrictions on deposition in Pickens Canyon after the flood of January 1, 1934. Their conclusions are fully supported by the present study, insofar as similar conditions existed in Arroyo Seco. Pickens Canyon, near La Canada, California, lies some 5 miles west of the mouth of Arroyo Seco Canyon. It is much smaller with a canyon section about 2 miles long. Jam deposits are not specifically mentioned, but the mechanics of transportation and deposition along the canyon are presented. The following sections attempt to apply this work to Arroyo Seco and to extend it to include the effects on the character of the sediments.

MECHANICS OF MOVEMENT OF FLOOD DEBRIS

The movement of sediment by running water has been studied by numerous workers, and from these investigations a number of universal principles have emerged. Several generalizations are used here as a basis for a qualitative statement of the problem. The ability of a stream to transport material may be expressed as its capacity (the quantity it is capable of carrying per unit time) or as its competence (the sizes it can move). In general the competence increases as the capacity increases, but the relation is not simple. Gilbert (1914) showed that capacity increases with the slope and with the discharge, as well as with the fineness of the debris.

Bed load is carried along the bottom of a stream, and suspended load at some distance above the bottom. Saltation occurs in an intermediate zone. Researches by Lane and Kalinske (1939) show that the distinction between bed load and suspended load is partly a function of the turbulence intensity near the bed, which controls the sizes, shapes, and densities of particles that may be lifted from the bed. The continued movement of material in suspension depends upon a concentration gradient of solid material in the vertical section (O'Brien, 1933), so that turbulent diffusion may carry more material upward than is carried downward by the same process. This excess upward diffusion is counterbalanced by the settling velocities of the particles themselves, and the interplay of these two factors establishes an equilibrium condition of suspension transport.

The movement of bed load is a function of bottom velocity, and recent work (Shields, 1936; Rubey, 1937) has shown a relationship between the thickness of the laminar sublayer in the flowing water and the sizes and amounts of material moved. Bed load movement can also be expressed in terms of discharge or of average velocity, which are known to be related to transportation, but perhaps not so fundamentally as is the bottom velocity. Nevertheless, any factor which changes the average velocity of the water will also change the capacity.

In floods the conditions of stream flow are much exaggerated due to the high discharge. A stream in flood may have a discharge a thousand times as great as the same stream at mean low water; in large floods this ratio may be even higher. Although conditions are thus much more violent during floods, there seems to be no valid reason why the same underlying physical laws of sediment behavior should not hold. Perhaps the distinction between bed and suspended load loses some of its sharpness. The steep gradients in mountain canyons and the high degree of turbulence generated by irregular surfaces throw into temporary suspension relatively large material. Equilibrium conditions, in which the loads become more or less distinct, may never be reached, so that the normal downward increase of particle size in a vertical section is confused by the intermingling of sand and pebbles at some distance above the bed. That such heterogeneity may have been present is indicated in part by the intimate mixture of coarse and fine material within the central sedimentation unit of both Arroyo Seco and San Gabriel canyons. Practically all size distributions are polymodal, suggesting a mixture of bed, saltation, and suspended load. Many exposures contain pebbles completely surrounded by sand, which renders it unlikely that the material represents two episodes of deposition, as an openwork gravel with the later infiltration of sand. Observations by Chawner (1934; 1935) also bear on this point. Splashings from the La Crescenta flood collected from roof drains showed on mechanical analysis a rather poorly sorted material with an appreciable content of pebbles as coarse as 8 mm. diameter. Apparently these splashings represent suspended material. There is little evidence of the relative importance of saltation and suspension in floods, but the high degree of turbulence in flood water suggests that if particles are freed from the bottom, turbulence intensity may be great enough to transfer them to the suspended load. The saltation zone in streams is usually rather shallow, whereas the main sedimentation unit in Arroyo Seco is locally several feet thick.

Such lines of evidence suggest that for some depth above the bottom the flood waters of Arroyo Seco carried a fairly concentrated load ranging from sand to moderate-sized pebbles. Toward the base of the zone cobbles and larger fragments rolled or slid along, behaving as bed material. The evidence points to simultaneous deposition from this concentrated zone, which would require a relatively abrupt velocity decrease to drop medium as well as coarse grains. No doubt the cessation of movement of coarse particles in such a load results in the trapping of some finer material, which may add to the content of sand and silt in the final deposit.

CONDITIONS CONTROLLING DEPOSITION

Figure 18 shows the plans and side elevations of two canyon sections; the upper has parallel sides and a uniform slope; and the lower has the same slope but contains a constriction. If water flows uniformly along the upper section, its depth remains constant, and the slope of its upper surface parallels the slope of the canyon bottom. Here sediment is transported at a given rate along the entire stretch. In the lower canyon the constriction completely changes the conditions. To maintain a given discharge the water in the constricted zone must move

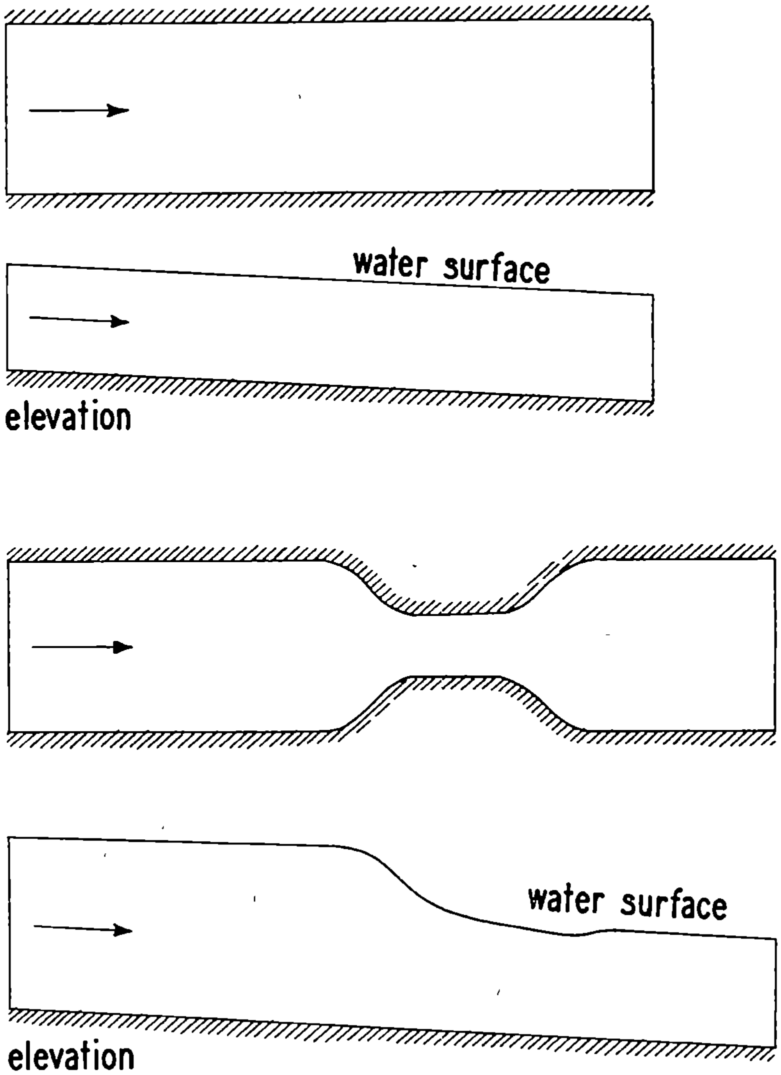

Figure 18.—*Plan and elevation of uniform and constricted channels*
Showing water profiles in steady flow

more rapidly. This can be accomplished only as the slope of the water surface steepens, inasmuch as the slope of the canyon is fixed. The result is that the water deepens appreciably at the upstream end of the constriction and thus backs up the water in the broader part of the canyon. The depth of water in the broad part decreases progressively upstream at a rate too slight to be shown in the figure, but over any considerable stretch average velocity changes. The net effect is that velocity decreases downstream until the narrows is reached, whereupon it increases rather markedly.

This analysis applies to mild slopes, in which the normal depth is greater than the critical depth (Rouse, 1938, p. 290 ff.). It also applies to steep slopes under the same conditions of depth. If the approaching flow is supercritical a hydraulic jump usually forms on the upstream side of the constriction. The jump may migrate upstream and thus develop the situation shown in Figure 18, with the jump left of the diagram. Whichever situation holds, it seems likely from the geological evidence that deeper water and lower velocities were developed upstream from the constrictions, thereby causing deposition, whereas within the constriction the velocity increased again and permitted scour.

A bend in the canyon associated with the constriction causes additional loss of energy. This effect may be slightly offset by an increase in bottom gradient through the narrows or along the bend, which will permit more rapid flow with less surface adjustment, so that the backwater effects may diminish. The course of Arroyo Seco Canyon rapidly alternates between straight stretches and abrupt bends, as well as relatively broad and narrow portions. At each change of direction or width opportunity arises for backwater effects. Debris cones from the tributaries, partially blocking the canyon, also have the effect of constrictions.

The debris cones built out from the tributaries are probably due mainly to the change of gradient between tributary and main canyon. Some of the small tributaries from the west (such as Agua Canyon) have slopes of 500 feet to the half mile, which is more than four times the average slope along the main canyon. Many of these short tributaries had road fills in their upper portions. During the flood this material was delivered to the tributary mouth where much of it came to rest. The larger volumes of flood water in the main canyon, already taxed with heavy loads, were unable to remove the material as it accumulated.

An example of deposition controlled by a constriction and a bend was given in connection with site 9 (Pl. 6, fig. 2). An example of the constriction caused by a debris cone from a tributary was given in connection with Brown Canyon (Fig. 8). In the light of the analysis (Fig. 18), backwater effects would have occurred upstream of Brown Canyon, with deposition in the zone of deepened water. After the surface slope was adjusted (combined with some adjustment of bed slope) the remaining load could be carried through with no deposition until the constriction was passed. Conditions upstream from Brown

Canyon substantiate this analysis, inasmuch as the main canyon there was so choked with debris that it had to be cleared to allow space for the stream and temporary road.

Log and boulder jams may arise in several manners. If logs are swung sidewise even in a broad part of the valley, the abrupt locking which results creates in effect a dam, with its immediate upstream response,—a slackening of velocity and a piling up of water. The partially cleared water pouring over the jam or around it can produce scour downstream. The coincidence of jams with bends and constrictions is apparently due to the much greater chance of logs or boulders locking. The net result is a complex deposit involving the elements of damming and constrictions. A hydraulic jump would favor the formation of jams by its abrupt slackening of velocity in the transition zone.

The presence of reverse size gradients in the jams is still not adequately explained. The only work bearing on this subject which has come to the writer's attention is an experiment by Straub (1940), in which natural river gravel was studied in a flume. When the load was added slightly faster than the current could carry it, the bottom slope was built up, and the coarsest particles were carried to the downstream end of the flume. Mechanical analysis of samples taken at intervals along the flume showed an increase in mean size from 2.5 mm. to 5.0 mm. over 24 feet. The original sediment had a mean size of 4.0 mm. Thus the finer material was left near the upstream end, and progressively coarser debris moved down the flume. According to Straub (1940, p. 183):

> "Depending upon the flow conditions and type of material, segregation upon deposition may take place either from fine to coarse (from one end of the channel to the other) or the reverse. Much additional study is required both in field and laboratory before any definite conclusions can be drawn as to the laws governing this occurrence."

Although the physical law underlying this situation is not known, apparently the controlling conditions were a thickening of the bed in a downstream direction as deposition progressed. If this experimental work may be applied to the present instance, very likely the lenslike deposits of the jams furnished the necessary conditions by building up most rapidly at the downstream end, thus decreasing the gradient and setting the stage for this type of size sorting.

The response of the sediments in terms of size and shape along the canyon as a whole conformed strikingly to the mechanics of their movement and deposition. The observations in Arroyo Seco indicated that in comparable sections of the deposits the average size and degree of sorting showed no systematic variations along the valley. Similarly, when the data are confined to one rock type, as the Lowe granodiorite, the average sphericity varies nonsystematically, whereas the roundness increases downstream. Inasmuch as size, shape, and density largely control the settling velocity, these attributes should reflect the dynamical conditions at the site of deposition. Roundness,—of secondary importance in settling velocity, but of primary importance as an index of abrasion—is a function of the wear on the pebble over the distance traveled. These same relations were noted in San Gabriel Canyon, from which it was inferred that essentially similar dynamical conditions were present at successive log jams; that is, the abruptness and order of magnitude of the velocity decrease was the same, because the size and shape response was the same.

In Arroyo Seco (and perhaps in San Gabriel Canyon) the samples include both jam material and alluvial plain deposits. The dynamic effect of constrictions and sharp bends on large masses of water moving moderately rapidly may be considerable, and responses in velocity adjustment may be rapid, especially if hydraulic jumps are formed. The frequent coincidence of log jams and bends argues that many of the alluvial plain deposits upstream of the bend may have formed under conditions nearly as vigorous as in the jams. Where constrictions cause deposition, the response of the sediments upstream is strictly dependent upon conditions at the constriction, because of the control of the water profile surface, and hence of the velocity, by the constriction itself. Therefore it may often be necessary in studying sediments to locate the control point in order to evaluate the results.

PARTICLE WEAR AND SELECTIVE TRANSPORTATION

The movement of debris along natural streams is accompanied by two basic processes going on simultaneously: selective transportation and particle wear. Selective transportation is reflected in the segregation of certain materials into bed and suspended load, with the simultaneous formation of a lag deposit. Selection proceeds largely on a basis of size, shape, and density. Highly spherical particles, for example, may outstrip less spherical ones in bed-load movement, whereas the less spherical particles outstrip more spherical ones in suspension transport. Wadell (1934) has discussed these effects in connection with his concept of sphericity. Simultaneously with movement of the particles, especially in bed-load transportation, size, shape, and roundness change as a result of particles rubbing or striking against each other. Marshall (1927) has shown that these effects may involve impact (breakage), abrasion (rubbing of particle against particle), or grinding (crushing of

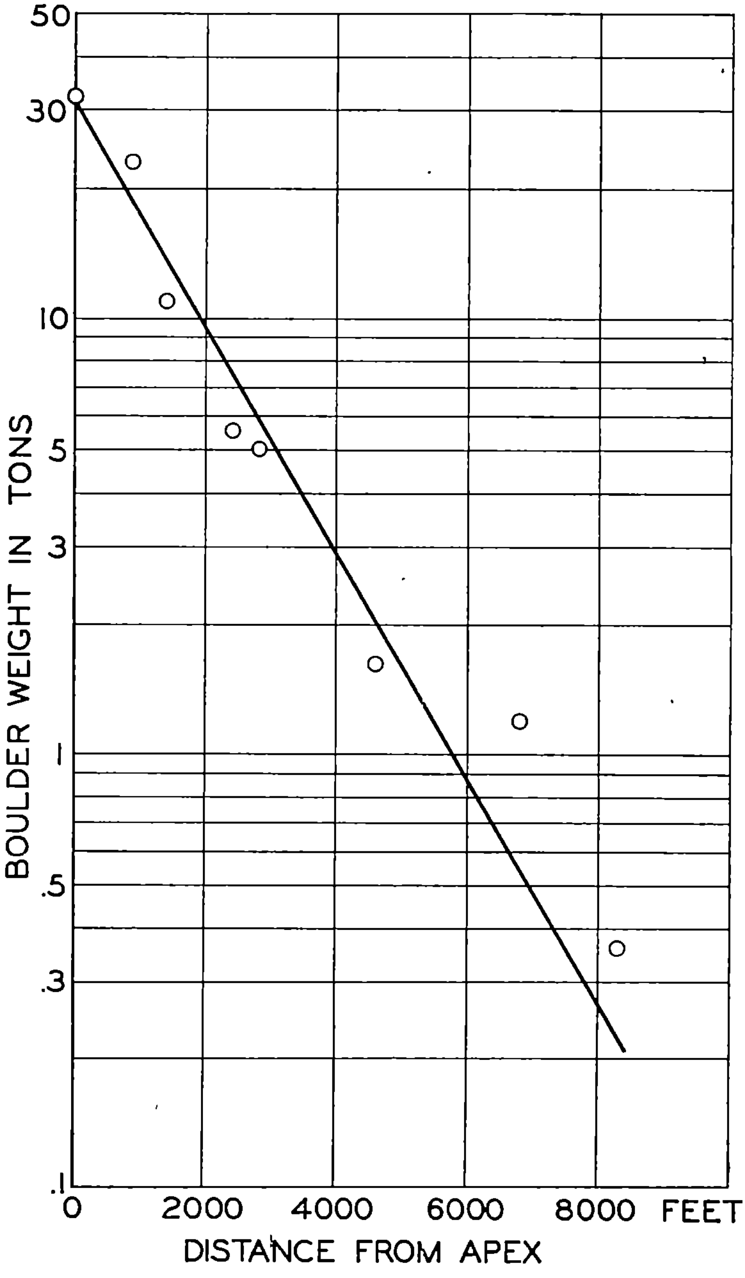

Figure 19.—*Variation in boulder size caused by La Crescenta flood, 1934*

(Data from Chawner, 1934)

small particles by large ones), all included in the term "wear". Which effect predominates depends upon the rigor of the process and the nature of the rocks. In general the average size of the particles decreases downstream over any considerable stretch. This is frequently attributed to wear, but it could also be due to lessened competence of the stream caused by a continually decreasing gradient. It is not possible, apparently, to segregate the two effects merely from the observed reduction in mean size; the problem is one of two unknowns and a single equation.

Under conditions of laboratory abrasion the size and shape change much less than the roundness (Krumbein, 1941b), so that considerable rounding may occur with negligible size and sphericity effects. In streams where abrasion is the only process of wear, apparently any rapid size change should be attributed to selective transport. Impact breakage has not been extensively studied, but the process undoubtedly results in a rapid decrease in size compared to abrasion alone. In steep canyons where greater rigor may permit impact breakage, the observed size decrease may actually be due to wear.

The relative effects of the two processes can be evaluated in the laboratory. Particle wear studies, in which the composition of the load remains constant, shed light on processes of abrasion and breakage independent of selective transport. Flume studies on the other hand permit an evaluation of selection where particle wear is entirely negligible. Rarely does only one effect occur in nature. The reversal of the size gradient in jam deposits can only be attributed to selective transport. Chawner (1934; 1935) measured a number of large boulders carried onto the alluvial fan during the La Crescenta flood of 1934. He tabulated the size of boulders in terms of their distance from the apex of the cone (Fig. 19). Despite the few observations, a marked size decrease is apparent over a short distance. The straight line indicates an exponential trend. The half-distance of the boulder size is found by determining the distance at which an ordinate of the straight line is half the value at its origin. This distance is 1100 feet from the canyon mouth. That is, boulder size decreases by half in the first 1100 feet across the fan. Here the short distance of transport, combined with the decrease in water velocity attendant upon its spread over the fan, indicates control by selective transportation.

Along Arroyo Seco the Lowe granodiorite boulders decrease markedly in size downstream. This effect is attributed to a statistical average of conditions over a long period of time, rather than just to one flood. Considering all rock types there was no over-all size gradient among the boulders along the canyon, due very likely to the continual influx of new material and to essentially similar conditions in the jams and other deposits. The only clue to possible effects of wear and selective transportation came from the data of Figure 11. The break in the straight line, which occurred near the zone between erosion and aggradation in the canyon, may reflect a change in the conditions of wear (as from breakage to abrasion); it may reflect an increase in the effects of selective transportation; or it may indicate a relative adjustment in both processes. The magnitude of the change is indicated by a ratio of the half distances of the two curves. In the steeper curve the half distance is 3.7 miles, and in the gentler curve it is 10.0 miles. Thus, on the average, the rate of apparent size decrease in the upper part of the canyon is more than 2.5 times that in the lower reaches.

In more detailed studies of wear and selection it may be necessary to include shape and roundness of the boulders. Angularity of the boulders over any considerable stretch may indicate continuous impact breakage, whereas progressive rounding may indicate less rapid abrasion. No detailed measurements were made, but angular boulders with evidence of fresh fracture were rare. It is the writer's present opinion that selective transport is on the whole more important than wear even during floods, especially in the lower part of the canyon.

SUMMARY AND CONCLUSIONS

The present study extends and amplifies the previous year's work on flood deposits. Certain tentative conclusions reached in San Gabriel Canyon were strengthened by evidence from Arroyo Seco; other inferences were modified by additional data. The combined studies furnish sufficient data for at least a qualitative reconstruction of the physical processes which operated during the flood.

Field work showed three main kinds of deposits formed by the flood: log and boulder jams at bends and constrictions; typical alluvial plain deposits in broad parts of the canyon; and alluvial cones at tributary mouths. These deposits were controlled by one or more of the following factors: abrupt decrease of gradient as at tributary mouths; abrupt velocity decrease associated with locking logs or boulders; velocity decrease caused by water profile adjustments at constrictions; and perhaps abrupt velocity decrease caused by hydraulic jumps.

Laboratory analysis of gravel samples showed no systematic variation in size or shape from corresponding parts of successive flood deposits, although in any single deposit (mainly jams) the size increased markedly in a downstream direction. Pebble roundness showed a systematic increase downstream along the entire valley. Pebble orientation showed uniform upstream imbrication even in the steep upper part of the canyon, and further studies in San Gabriel Canyon showed that reverse imbrication found there was due to local causes.

A limited amount of data were presented on the relative effects of boulder wear and selective transportation along the canyon. These data suggest a more rapid apparent size reduction in the upper reaches of the canyon, but they do not permit an evaluation of the direct cause of the size decrease.

It became increasingly evident during the study that the basic unit for sedimentary investigations is not the sample, but the entire depositional environment. The analysis of sediments should be integrated with physical principles of particle transportation and wear, but such integration is at present hindered by lack of complete data on flood discharge, sediment load, and velocity changes along the canyon. The superposition of large flood phenomena on more normal processes of stream work also introduces a complicating factor. It is believed, however, that the geological approach offers the most direct means of reconstructing flood conditions until such time as more direct measurements of flood phenomena are available.

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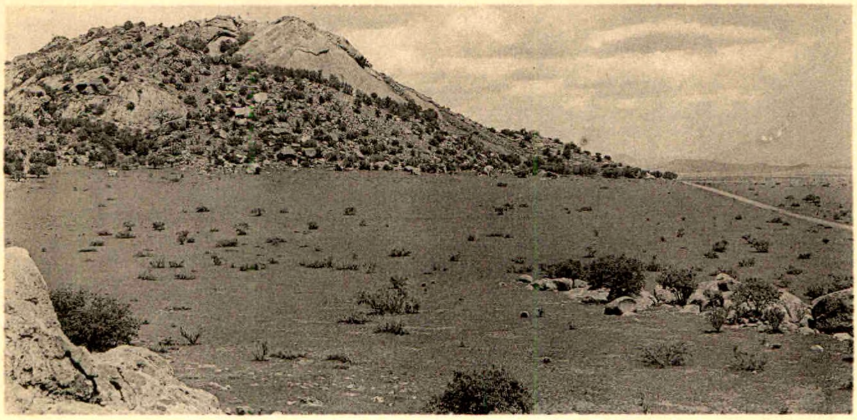

FIGURE 1. CONICAL HILL OF SYENITE, LOCALITY 1

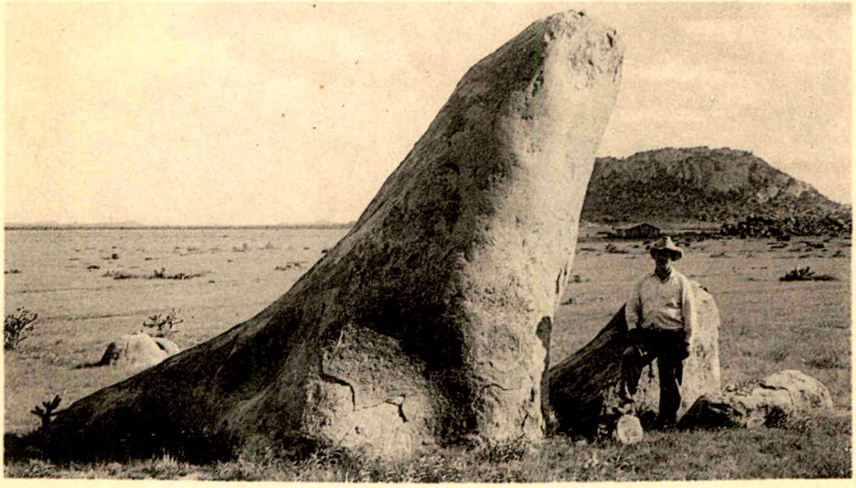

FIGURE 2. SYENITE BOSS RETAINING POLISH ON UPPER PART, LOCALITY 1

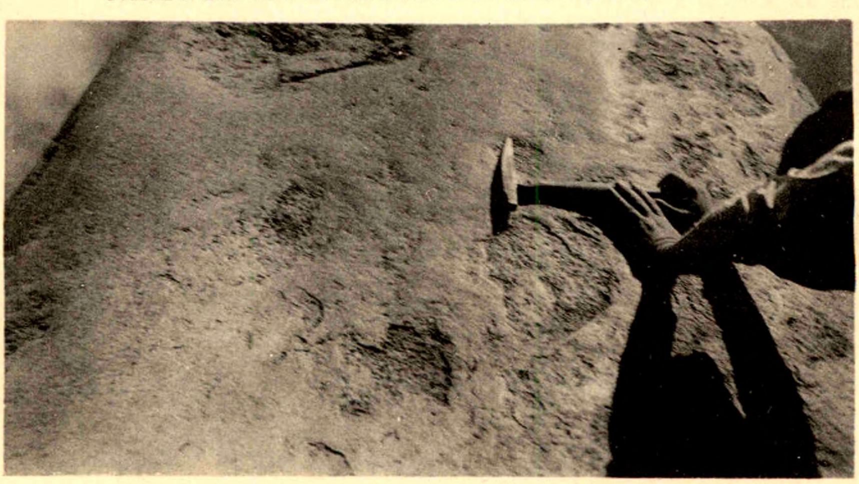

FIGURE 3. DETAILS OF POLISHED SURFACE ON [SAME] ROCK SHOWN IN FIGURE 2

WIND–POLISHED ROCKS IN THE DAVIS MOUNTAINS

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WIND-POLISHED ROCKS IN THE TRANS-PECOS REGION, TEXAS AND NEW MEXICO

BY KIRK BRYAN AND CLAUDE C. ALBRITTON, JR.

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ABSTRACT

Large surfaces polished by wind-borne dust on bedrock and boulders are abundant along the margins of the Davis, Quitman, Finlay, and other mountains in the Trans-Pecos region. At present the polished surfaces are being destroyed by weathering, and new polishing is not taking place. The polished surfaces are thus relics of a past time when wind-borne dust was a more effective agent of corrasion than at present.

Indian petroglyphs engraved on and partially destroying wind-polished surfaces in the Finlay Mountains are unmodified by later wind action. Near-by middens contain fragments of pottery which, according to the archeological chronology, were made between 1200 and 1300 A. D. Granting that the associated petroglyphs and pottery are contemporaneous, the picture-writings are 600 or 700 years old, and the wind-polished surfaces are still older.

Alternation of relatively humid and relatively arid intervals during the late Quaternary has already been postulated on evidence provided by alluvial deposits in the Davis Mountain area. It is suggested that the wind-worn surfaces here described were formed during one of the dry periods. The longest and most important such interval intervened between deposition of the Neville and Calamity formations. The polished surfaces presumably date back to this dry period which is to be correlated with the so-called Postglacial Optimum 5000 to 7500 years ago.

INTRODUCTION

In 1891 Johannes Walther, traveling through Trans-Pecos Texas, found numerous polished boulders which he later mentioned in *Das Gesetz der Wüstenbildung* (1900, p. 101-102). Although Walther was not specific as to localities, he apparently refers to certain wind-abraded rock surfaces in the northern Quitman Mountains, possibly the same examples as can be seen today along U. S. Highway 80 about 10 miles west of Sierra Blanca (Fig. 1, locality 4). He attributed the polish to the sand-blast that scours the dry basins and foothills between infrequent rains—hence to a process intermittently active at present.

Investigations in western Texas made by the authors during the field seasons 1938 and 1940 provided opportunities for examining polished rock surfaces at six localities. The polish is everywhere being destroyed by weathering and is thus a relic of a time when the wind was locally a more effective agent of abrasion. This paper deals with the distribution and forms of the wind-worn surfaces and with the problems of their significance and antiquity.

Field work for this paper was begun in 1938 when the writers, aided by a grant from the Penrose Bequest of The Geological Society of America, were studying Quaternary deposits in the Davis Mountain area, and has been extended by Albritton while engaged in field work devoted largely to other problems.

DESCRIPTION OF WIND-POLISHED ROCKS

DISTRIBUTION

Localities at which wind-polished rock surfaces were observed are indicated by numbers 1 through 6 on Figure 1. The first three are along the borders of the Davis Mountains, where the annual rainfall averages less than 17 inches. Localities 4 to 8 are along or near the margin of the Hueco Basin, in a drier region with an annual precipitation averaging around 8 inches. Locality 4 is at the northern end of the Quitman Mountains; 5 and 6 are near the southeastern and northwestern ends of the Finlay Mountains, respectively. Lang (1941) has described polished rock surfaces in the Cornudas Mountains (locality 7) and Hueco Mountains (locality 8) which may be similar in origin to those observed farther to the south and east.

DAVIS MOUNTAINS

Locality 1, 12 miles southwest of Fort Davis, may be reached by the Valentine-Fort Davis road. This locality is in a transitional belt between the Davis Mountains on the north and a broad pediment that slopes toward Alamito Creek to the South (Albritton and Bryan, 1939,

p. 1451-1452, pl. 1). Northern extensions of the pediment indent the mountain front, and scattered steep-sided residuals rise a few hundred feet above the plains to the south and west. Three such outlying hills lie along a meridional line connecting the Finlay and Merrill ranch houses. (*See* Valentine topographic sheet.) Each is roughly conical and composed of fine- to coarse-grained syenite. Joints divide the rock into large blocks, which tend on weathering to assume rounded shapes as a result of exfoliation. The peripheries of the hills are studded with these bosses and boulders (Pl. 1, fig. 1).

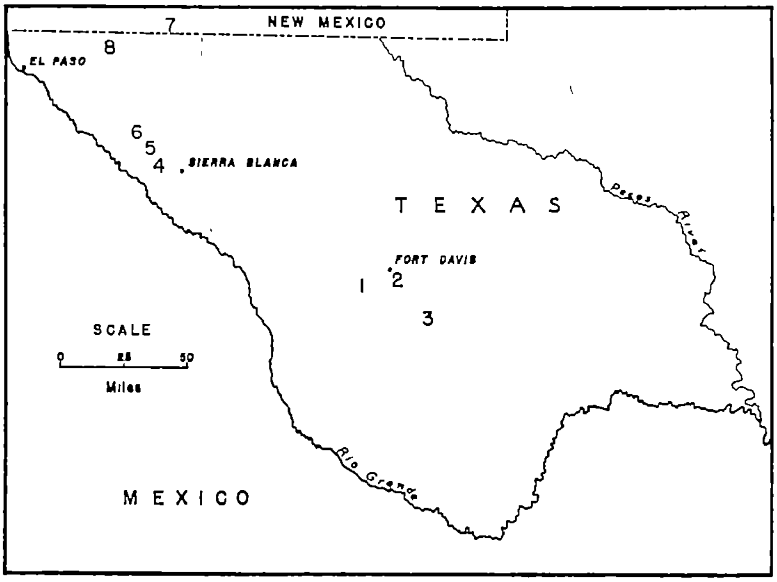

FIGURE 1—*Map of Trans-Pecos Texas showing localities of wind-polished rock*

Dozens of wind-worn rocks occur along the bases of the three hills. One striking example (Pl. 1, fig. 2) has a smooth and polished surface that contrasts strongly with the rough granular exterior which is the ordinary result of weathering on this rock. The polish is mostly on the feldspar crystals. Under a magnification of 10 times or greater, numerous small scratches trend at random. These apparently smooth surfaces resulted from abrasion by an agent capable of producing fine rather than coarse scratches or chatter marks. The ferromagnesian minerals are generally marked by dull surfaces which consist of small pits, most of which seem to have developed by differential weathering after the polishing. Near the base the blocks are being actively exfoliated in thin shells, so that the polished surface has been or is being carried off on the shells of rock. Near the top, however, are older exfoliation scars, some of which were nearly effaced during the scouring which formed the polish (Pl. 1, fig. 3). Thus the contours of this boss were being modified by exfoliation before and during the polishing. Active exfoliation during the period of wind abrasion may in part explain the absence of flutes or facets.

Smooth and lustrous surfaces like those described above are present only on rock surfaces that are relatively sound and unweathered. When the surface is deeply weathered and crumbly, the polish is duller and is limited to the firmer feldspar grains standing in relief above the irregularly eroded depressions of more decomposed zones.

Polish occurs only on rocks around the bases of the hills and on upstanding residuals away from the bases. The vertical limit of the polish is about 12 feet above the local pediment level. Rounded boulders generally have a greater proportion of polished surface than do angular boulders, on which the polish is usually restricted to exposed corners and edges. Rocks less than 50 yards from the base of the nearest hill generally are not polished on the sides facing the hill; those farther away may be polished on all sides. The same effect is seen on a larger scale in the distribution of wind-worn rock surfaces around the three hills; the most isolated hill lying the greatest distance south of the Davis Mountain front and thus most exposed to winds from all directions has wind-worn rock surfaces all around its base, whereas the northernmost hill nearest the mountain mass has no wind-worn rock surfaces on the northeast.

The massive syenite of these hills seems to have taken and also to have preserved polish most successfully. The location of hills of other rock types is in all respects similar, and presumably wind action has also occurred, but the evidence has been lost by more rapid weathering. At present the wind apparently is not corroding hard rocks at this locality. There are no active dunes here or near by, and the smooth pediment floor with its thin veneer of alluvium in transit is effectively protected against deflation by a carpet of grass (Pl. 1, fig. 1).

The absence of dunes or of patches of wind-transported sand or dust either near these semidetached hills or on the mountain slopes to the northeast is a remarkable fact. It must be attributed to extensive erosion by rain wash and stream action since the time the polish was formed on the rocks.

At Locality 2 on the north side of the Fort Davis-Alpine highway 4 miles southeast of Fort Davis a layer of dense gray trachyte crops out as a low bluff trending northeast and facing the broad headwater valley of Limpia Creek to the north, west, and southwest. (*See* Fort Davis topographic sheet.) The bluff is thus exposed to all but southeast winds. Edges and corners of joint blocks are polished up to 6 feet above ground level. In spite of the drab color of the rock, the polish is remarkably perfected on sound rock. On weathered rock the polish is duller. The polished surfaces are smooth, passing without interruption over glassy sanidine phenocrysts and the gray aphanitic matrix alike. Weathering subsequent to polishing has loosened many feldspar crystals, so that the surfaces are commonly indented with angular, unpolished pits.

Locality 3 is on the north side of the Alpine-Terlingua road, 3 miles southeast of Alpine. Wind-worn rocks are scattered over the crest of a low, grassy divide between two headwater branches of Lencito Draw and exposed to wind from all directions. (*See* Alpine topographic sheet.) Bedrock is a microcrystalline brownish vesicular latite (?), the vesicles of which are filled to varying degrees with chalcedony and drusy quartz. The volcanic rock juts out in low angular hillocks, and angular boulders of various sizes are scattered over the divide. Effects of wind abrasion are restricted to projecting edges and corners, and though no larger surfaces are noticeably worn the polish where developed is lustrous. None of these rocks rise more than 2 feet above the surrounding grassy surface.

QUITMAN MOUNTAINS

Wind-worn rocks are abundant in the northern Quitman Mountains (Fig. 1, locality 4). Examples can be seen south and southeast of U. S. Highway 80 at the bend around the northern end of the mountains about 10 miles west of the town of Sierra Blanca. They are restricted to a narrow belt along the base of the mountains, where foothills and spurs meet the almost featureless surface of a pediment that extends far to the south and west over the Hueco Basin. (*See* Fort Hancock topographic sheet.) Bedrock is syenite, and the worn surfaces exhibit the same characteristics of distribution, configuration, and preservation already described for similar rocks at locality 1. Polish is best developed on corners, edges, and sides of boulders facing the west; thus it appears that the abrading winds came from the west, and therefore from the direction of the Hueco Basin.

The sandy surface of this basin is inadequately protected against deflation by greasewood and other bushy plants. The prevailing southwesterly winds therefore produce sandstorms, but the intermittent sand blast of the present time is unable to maintain polish on the rocks. The once smooth and lustrous surfaces have been so nearly destroyed by exfoliation that they are now rough over most of their surfaces and retain only a few projecting scales with polish. J. Walther (1900, p. 29) noted this fact, but attributed the exfoliation to torrential rains that fell a few days before his arrival. The writers have observed the same examples of polished boulders before and after numerous rains without being able to detect any change in the surfaces. Destruction of polished surfaces in this locality has progressed more slowly than Walther supposed, and if the writers' estimate on antiquity of the polish is correct the destructive effects have resulted from centuries of weathering.

Mr. R. M. Huffington, who is preparing an account of the geology of the northern Quitman Mountains, reports that wind-worn stones are widely spread along the peripheries of the range at many localities other than the one here described.

FINLAY MOUNTAINS

Locality 5 is a small hill rising a few feet above the level of the Hueco Basin at the southeastern end of the Finlay Mountains. Boulders and residuals of dense gray hornblende porphyry are polished. The hill is exposed to wind from all directions. As the wind-worn surfaces are like those of the Quitman Mountains, no further description is needed.

Locality 6 marks the site of the Wilkie ranch house, situated on the west bank of Campagrande Draw in the western Finlay Mountains (Lat. 31° 23′ N., Long. 105° 39′ W.—Fort Hancock topographic sheet). The corners and edges of sandstone blocks are polished as high as 12 feet above the ground. The blocks are derived from near-by outcrops of the massive sandstones of the Cretaceous Cox formation. Concentration of iron compounds resulting from the decomposition of marcasite nodules within the body of the rock stains the surfaces of the blocks a reddish brown. The outer rinds, a fraction of an inch thick, are more indurated than the normal granular sandstone. Wind polish is superimposed on the protective rinds. The polished rind of these rocks is, according to Hume (1925, p. 211), the true desert varnish although many authors consider that the presence of the rind alone without superimposed wind polish adequately represents desert varnish. The rock surfaces are exposed to winds sweeping along or across the broad Campagrande Valley from the north, south, or east. After camping at this locality, one can attest that strong winds transport sand and dust in disagreeable quantities. Nevertheless, these storms are inadequate to replace the polish which is being destroyed by weathering.

HUECO AND CORNUDAS MOUNTAINS

Lang (1941) has recently described polished rocks in these mountains. Like those of the Finlay-Quitman area, these localities lie east of the broad Hueco Basin (Fig. 1, localities 7 and 8). The rocks affected are granite, and the polished surfaces were found by him on the upper walls of "crevice caves" and niches, and on large boulders on the "surrounding apron of detrital wash". Polish is mostly on the south side and extends from points about 2 feet above ground to 7 or 9 feet above ground. Lang attributes the lack of polish in the zone up to 2 feet above ground to a recently lowered ground surface. The writers' observations at other localities indicate that exfoliation is more complete near the ground, and it is thus reasonable to conclude that the polished surfaces described by Lang are now being destroyed by exfoliation as are others in the same region.

POLISH BY WIND-DRIVEN DUST

Polished rocks widely distributed over Trans-Pecos Texas are numerous in the Davis, Quitman, and Finlay mountains and are reported from the Hueco and Cornudas mountains. Doubtless other localities will be discovered. The sporadic distribution is due to lack of exploration and also to the existence, in these localities, of rocks particularly suitable for the preservation of the polish.

Fluted or grooved surfaces have not been found, nor are near-by deposits of wind-blown sand known.

Although polish is developed on the surfaces of different types at different localities, four characteristics are common to all examples: (1) The polish is restricted to upstanding boulders and to bosses of bedrock situated along the borders of residual hills and mountains adjacent to surrounding pediments. The localities are exposed to winds from several directions. (2) The polish is generally restricted to surfaces 12 feet or less above the local ground level. It is more marked on the southern and southwestern sides, but locally appears on all sides of a given boulder. (3) The polish is superimposed on surfaces which owe their form primarily to weathering and not noticeably to fluting and faceting by sand blast. Weathering is destroying the polished surfaces. Where the polish has been removed, as by exfoliation, there is no evidence that it is being restored.

Fluting and carving of rocks, described by many authors (Blake, 1855; Gilbert, 1875, p. 82-84; Cloos, 1911, Pl. 2, fig. 9; Hume, 1925, p. 64; Blackwelder, 1929, p. 256; Russell, 1932, p. 101-103; Denny, 1941, p. 256-259), seem to be related to the movement of coarse sand by winds of fairly constant direction. Polishing of bedrock and pebbles also occurs in many of the same localities. It is produced by the impact of finer grains such as dust. Direct reference to polishing is made by many authors, and, recently, rocks polished and pitted in Pliocene time have been described and illustrated by Smith (1940).

In order that these rocks and surfaces here described be polished, it is necessary that the agent should act on all sides of isolated boulders at heights up to 12 feet above the ground. This distribution of polish and the fineness of the scratches of which it is composed imply that the dust was swirled in the irregular eddies and vortices of the wind. The winds were presumably variable in direction throughout the year, or boulders could not have been polished on all sides. The greater number of polished surfaces and general tendency for more polish on the southwest sides of hills, as at Locality 1, testifies to prevailing winds from the southwest similar in direction to those of the present.

Lang (1941) has raised again the question of the role of animals in the production of polished surfaces. Undoubtedly the sand-impregnated coats of animals—particularly the grazing animals—are capable of producing a polish in which the presence of the oils and waxes of the skin is a factor. Bryan (1925) has observed at Tinajas Altas in the Papago Country of Arizona polished surfaces at the rubbing places and lairs which mountain sheep use in the heat of the day. As Lang states: "The smooth and highly polished surfaces formed on posts, pipes and other hard and resistant objects, produced by cattle satisfying their urge to scratch, is a common sight about water holes in the western ranges."

Schoewe (1932) explains the "mirror polish" found on only one side (northwest) of an isolated erratic in Brown County, Kansas, as due to rubbing by bison. This boulder differs from others in this part of Kansas in having a much higher or "mirror" polish, in being located near water, and in being the only shelter or rubbing place for miles around. Other boulders in this part of Kansas have only minor areas of polish, and Schoewe implies that the polish on these boulders may be attributed to wind action accompanying the last glaciation. Thus the one boulder in Brown County is a special case.

K. Walther (1924) has described the polished surfaces of isolated granite boulders on the plains of Uruguay which he attributes to rubbing by sheep. However, in 1930 he retracted this explanation in favor of the formation by chemical weathering of a siliceous surface induced by a process of "case-hardening." He does not, however, wholly abandon the idea of animal polishing which may apply in special instances.

In the Trans-Pecos region there are so many of these surfaces, some so close to the ground, some so high above the ground, some in such relatively protected positions, that polishing by animals seems unlikely. One must suppose an indefatigable perseverance in rubbing by large numbers of animals. Such industry is difficult to imagine in a state of Nature. Herds of cattle or sheep furnish no true pattern of behavior. The provision of watering places by man has vastly increased the carrying capacity of the range so that domestic cattle are much more numerous than the herbivorous animals of the past could have been. Furthermore, they follow their instinct to loiter in the heat of the day without fear of the larger carnivores. They thus have greater leisure for indulging in rubbing on upstanding objects.

One cannot affirm that rubbing by animals may not have caused a polish or increased the polish on any one rock or boulder that may have been especially well-located with regard to a natural watering place. The numerous and widely scattered polished surfaces of the Trans-Pecos region have a distribution unrelated to natural watering places. Some are polished to a height above the ground of 12 feet or too high for animal polishing. The number of surfaces and their distribution require an efficient and widespread cause such as wind-driven fine sand or dust. That none of this dust remains in the vicinity of the surfaces is an obstacle to the acceptance of this explanation. However, periods in the past when dust was moved by the wind have been established, and intervening time has been long enough to provide opportunity for the removal by erosion of the dust near the rock surfaces.

ANTIQUITY OF THE POLISHED SURFACES

Locality 5 is also the site of a former Indian encampment that is remarkable for a number of petroglyphs consisting of the outlines of animals and various conventional symbols (Osburn, 1941; Howard, 1941). All the drawings are pecked into the sandstone boulders, and some are cut into the wind-worn surfaces already described. No new polish has formed where drawings are found. Hence the wind polish is older than the petroglyphs. The ground surface around these inscribed boulders is littered with flint chips and other refuse of ancient Indian camps. From these middens pottery fragments were obtained which H. P. Mera attributes to people who lived between 1200 and 1300 A. D. (Campbell, *et al.*, 1941). As the camps and petroglyphs appear to be contemporaneous, the polished surfaces are more than 600 to 700 years old.

If the polished rocks are attributed to some past, relatively more arid time, the sequence of Quaternary events already known for the Davis Mountain area offers two possibilities for correlation. In 1939 the writers showed that Quaternary alluvium in the valley flats adjacent to the Davis Mountains is divisible into three formations: Neville, Calamity, and Kokernot—in order from oldest to youngest. Disconformities separate the formations, providing that the three stages of alluviation alternated with stages involving cutting of gullies or arroyos. The stages of alluviation appear to have been relatively humid times, whereas those of arroyo-cutting were relatively dry. Probably the wind was an active agent of deflation and corrasion both during the Neville-Calamity interval and the more recent Calamity-Kokernot interval.

Wind work in Neville-Calamity interval is well established. Wind-blown dust in the caves and shelters of the Davis Mountains indicates that the grass-covered intramontane valleys of these mountains have at times been sear and windswept. The age of this dust has been discussed by Kelley (Kelley and Campbell, 1940). In Rock Cave, for instance, excavation showed an upper cultural layer containing artifacts of the Chisos focus overlying several feet of angular rock fragments derived from the roof. The rock fragments, in turn, rest on wind-borne dust. In another shelter, known as Alpine 2:7, debris containing artifacts of both the Livermore focus and the Chisos focus overlay debris containing artifacts of the Pecos River focus which in turn rested on fine reddish-yellow dust. In both these caves the dust is sterile, but in Hord Rock shelter Smith (1934) found bones of a horse and a musk-oxlike animal at the top of the dust layer. As artifacts of the Pecos River focus occur in the upper part of the Calamity formation, the dust in the caves is presumably older than that formation. This conclusion is confirmed by the fact that the bones of extinct animals do not occur in the Calamity. Hence in 1938 the writers assumed that the dust and the rock polish both belong to the interval of erosion marked by the disconformity separating the Neville and Calamity formations.

Kelley and Campbell (1940, p. 158-162) discuss at length the age of the cultures involved in the stratigraphy of the Davis Mountain area without reaching a conclusion as to definite dates in years. "For the present, therefore, no definite dating of the Calamity formation and its equivalents, or the immediately preceding and following erosional periods, or the included cultural remains is practicable."

It is possible, however, to speculate on the antiquity of the Neville-Calamity erosion period. The Neville formation contains bones of an extinct horse and also an elephant. The presence of an elephant implies that a more luxuriant vegetation and a wetter climate also existed. Physical evidences also imply climatic conditions wetter than the present. Thus it is inferred that the Neville dates back to the wet period coincident with one of the readvances of the retreating continental glaciers. The last advance, recognized at Cochrane, Ontario, and in the Fennoscandian moraine of Sweden and Finland, has been dated at 10,000 years ago. (*See* review of the problem, with bibliography, *in* Bryan and Ray, 1940.) The Neville is at least as old as this date and might be earlier. The ensuing Neville-Calamity erosion interval was warmer and dryer than the present and might coincide with the warm period known in Europe as the Postglacial Optimum. The best date for this interval appears to be that of Fromm (1940) who fixed it in the years 5000 to 7500 before our time.

It should be remembered also that the salt lakes of the Great Basin area were dry at a time subsequent to the last glaciation, definitely marking a period of aridity in the post-glacial of western United States. Its date has been estimated by Van Winkle (1914) and others by computing the present salt content of the lakes and dividing this quantity by the present annual increment of salt added by the inflow of river water. The resultant figure makes the present salt lakes 4000 years old, but that is obviously too small a figure. In a gradual amelioration from a previously dry period to the present mean rainfall, the inflow from the rivers both of water and salt would not begin suddenly at the present amount, but should gradually increase. Thus the divisor in Van Winkle's calculation should be smaller and the length of time longer. Further, there have been minor fluctuations of climate in the interval with an unknown effect on the inflow of salt. Nevertheless, the figure of 4000 years back to a dry time is of the same order as the figure of 5000 years back to the warm period in Europe. It is a fair presumption that these fluctuations toward the warm in Europe and toward the dry in western America are contemporary.

If the Neville-Calamity interval of erosion and relative dryness falls in the great postglacial warm period, the Climatic Optimum of Europe, the rocks were polished more than 5000 years ago.

The possibility must be considered that the polished surfaces date from the Calamity-Kokernot interval of erosion and relative dryness. Kelley and Campbell (1940, p. 64-71, 160-162) discuss the date of this interval. Artifacts of the Livermore focus raveled from a village refuse heap into the post-Calamity arroyo at Nine Point Mesa. As a typical Livermore projectile point has been found at Mogollon Village, New Mexico, in an archaeological horizon dated by tree rings as ending about 900 A. D., it can be postulated that the arroyo may have been cut at least as early as this date. In the Kokernot formation the known artifacts are given a date that may be as early as 1200 A. D. However, this dating is uncertain. It is apparent that we are trying to cut fine distinctions with too dull a knife. Kelley and Campbell admit that the beginnings of Kokernot deposition may have been only a short time prior to 1400 A. D. and also that the post-Calamity erosional period may have been coincident with the great drought of the late thirteenth century recognized in tree rings. In whatever way these difficulties in dating may be resolved eventually, the Calamity-Kokernot interval was short and lasted at most 100 to 300 years. It occurred not more than 1000 nor less than 500 years ago.

There is no local evidence of wind movement in the Calamity-Kokernot interval, although greater activity than at present may be presumed. If the surfaces were polished in this interval, their subsequent preservation is no great problem. If they were polished in the Neville-Calamity interval, however, the surfaces must have persisted for 5000 years or more. One might suppose that the rocks at localities adjacent to the Hueco Basin may have been polished during the first dry period and repolished in the second, but this supposition seems less probable in the Davis Mountain localities. The caves in the near-by Davis Mountains received much dust in the first interval but little or none in the second, indicating that the Calamity-Kokernot interval was not sufficiently dry or prolonged for effective wind action in this area.

The writers hold that the balance of evidence indicates that the polishing occurred in the Neville-Calamity period of erosion and wind action, about 5000 years or more ago. This conclusion involves a persistence of the polish in spite of the intervening 5000 years of weathering. The persistence of the polish testifies to the relative resistance of the rocks involved and to the slow weathering of arid regions.

Wind-carved blocks of gneiss many of which retain a high polish are found on the Home moraine in the Cache la Poudre Valley of Colorado (Bryan and Ray, 1940). These polished rock surfaces may have been covered by wind-blown sand, of which patches remain, for part of the time since the surfaces were wind-scoured. The wind-action goes back to middle Wisconsin time. Similar wind-cut and wind-polished rock fragments occur on the Corral Creek moraine. There is little evidence that they have ever been protected by a cover of sand. These stones were polished by the winds that blew off the ice as it retreated after building the moraine some 25,000 years ago. Therefore, sufficiently resistant rocks in suitable climates will retain a wind polish for many thousands of years.

There is no direct evidence for the contemporaneity of all the polished surfaces of the Trans-Pecos area except, that from place to place, there is the same degree of preservation. Therefore, some of the surfaces, particularly those near the Hueco Basin, may have been polished or at least repolished. The appropriate time is the Calamity-Kokernot interval which has a date ranging from 900 A. D. to 1400 A. D. or 500 to 1000 years ago.

SUMMARY

Polished rocks at numerous localities in the Trans-Pecos region and mostly in Texas resulted from past wind action. The number and magnitude of the surfaces are too large for formation by the relatively haphazard process of polish by animals. The wind polish is now being destroyed by weathering. Although there is no direct evidence that the surfaces were abraded contemporaneously, the inference from their common state of preservation is clear. The available evidence indicates that the polish was formed in the Neville-Calamity period of erosion and relative dryness. This interval is presumably the great period of warmth and, in western United States, of aridity 5000 to 7500 years ago. Such a conclusion requires the preservation of the polish for a period of at least 5000 years.

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LATE JURASSIC FOSSILS FROM CUBA AND THEIR ECONOMIC SIGNIFICANCE

BY RALPH W. IMLAY

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ABSTRACT

The Viñales limestone fauna is correlated with the upper Portlandian on the basis of the ranges of its normally coiled ammonites and their similarity with species in the upper Portlandian of Mexico and Argentina. Genera not known below the upper Portlandian include *Corongoceras, Micracanthoceras, Durangites, Lytohoplites,* and *Parodontoceras*. Genera not known above the Portlandian include *Metahaploceras, Physodoceras,* and *Durangites*. Genera not known above the lower Tithonian include *Corongoceras* and *Lytohoplites*.

Previous correlation with the Lower Cretaceous was based mainly on identity of aptychi and similarity of uncoiled ammonites to species in the Lower Cretaceous of southern Europe. Association of these forms with normal ammonites permits no doubt as to their Portlandian age and suggests that aptychi must be used with caution in detailed stratigraphic work. Nothing in the fauna suggests the Kimmeridgian or Tithonian stages The much younger age of the fauna than that of the Argovian-lower Kimmeridgian ammonites, described by Sanchez Roig and Marjorie O'Connell, shows that the older fossils are from the San Cayetano formation beneath the Viñales limestone, and that these formations are separated by an unconformity involving parts of the Kimmeridgian and Portlandian.

Furthermore, the study shows that undoubted Lower Cretaceous rocks have not yet been found in the West Indies, that metamorphic rocks elsewhere in the West Indies may be Jurassic, and that the regions bordering the Gulf of Mexico have a similar Jurassic history. It suggests that the Palisade disturbance is late Jurassic.

Application of these conclusions toward forming a clearer picture of events in Jurassic and Lower Cretaceous times should aid in showing the time of origin and possible distribution of mineral deposits and should be particularly useful in explorations for petroleum in the deeply buried rocks of the Gulf region.

INTRODUCTION

The present study was begun with a fair degree of confidence that the Viñales limestone fauna was early Lower Cretaceous. Interest was first aroused by examining a small collection of ammonites and aptychi from Camagüey Province. Fossils from the Viñales limestone have been received periodically by the Geological Survey since 1929 and have usually been identified as probably Lower Cretaceous. This age assignment seemed reasonable on the basis of the similarities of the uncoiled ammonites and the aptychi to Lower Cretaceous Eurasian fossils and the rarity of uncoiled ammonites in the Upper Jurassic. Further confirmation of Lower Cretaceous age was furnished by several recently published statements, although nowhere had undoubted proof been published of rocks of Lower Cretaceous age in Cuba or in any other of the West Indian islands. A Lower Cretaceous age was apparently contradicted, however, by the supposed occurrence in the Viñales limestone of ammonites of Argovian-lower Kimmeridgian age described by Sanchez Riog and Marjorie O'Connell in 1920 and 1921. Subsequently, Dickerson and Butt (1935), on the basis of considerable field work, claimed that these Jurassic fossils actually were derived from the phyllites of the San Cayetano formation but were mixed locally with float from the overlying Viñales limestone. Furthermore, they contended that the Viñales limestone was separated from the San Cayetano formation by an angular unconformity involving much of Upper Jurassic time. As considerable confusion resulted from these contradictory statements of age and relationship, the writer decided that a published description of the common species of the Viñales limestone would form a sure basis for its age determination, would show the correct sequence of events in Cuba during Upper Jurassic and Lower Cretaceous and would afford more accurate correlations with rocks of similar age in other lands bordering the Gulf of Mexico and the Caribbean.

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The fossils were collected almost entirely by Roy E. Dickerson, R. H. Palmer, W. H. Butt, Emilio Alemán, Antonio Martínez, and N. E. Weisbord, during the years 1929 and 1934 while in the employ of the Atlantic Refining Company. Most of the collections from Santa Clara Province were made by Palmer, and most of those from Pinar del Río Province by Dickerson. One locality in northern Santa Clara Province was recollected by Emilio Alemán, D. W. Gravell, and John Klecker in 1941. A small but important collection from Camagüey Province was made by Thomas P. Thayer of the Federal Geological Survey in 1941. Particular thanks are due Doctor Dickerson for obtaining permission from the Atlantic Refining Company of Cuba to study their extensive collections, and for his careful checking of locality and stratigraphic data. The splendid co-operation shown by petroleum geologists in this, as in past studies, is deeply appreciated.

VIÑALES LIMESTONE

DISTRIBUTION AND THICKNESS

The surface extent of the Viñales limestone is shown on a generalized geologic map of Cuba compiled by Lewis (1932, opposite p. 534), and on more detailed maps prepared by members of an expedition from the University of Utrecht (Rutten, 1936, p. 1-59, 3 pls.; Thiadens, 1937, p. 1-69, 3 pls.; Vermunt, 1937, p. 1-60, 3 pls.; MacGillavry, 1937, 169 p.). The formation extends (Palmer, 1938, p. 20) about 375 miles from westernmost Pinar del Río Province to eastern Camagüey Province and crops out mainly in the northern part of the Island. Dickerson (1937, p. 418) notes that the formation:

> "is exposed in the Organ Mountains, Pinal del Río Province; in several small outcrops in northern Matanzas Province; in a worn-down cordillera 120 miles long in northern Santa Clara Province and westernmost Camagüey Province; and in a small area in northeastern Camagüey Province."—and—"probably underlies northern Havana Province as well."

The subsurface extent of the Viñales limestone toward the south is unknown, but Dickerson (personal communication) thinks that it is probably absent in southern Cuba as indicated: (1) by its absence on the Isle of Pines; and (2) the lack of seepages in the southern part of Cuba.

The thickness of the Viñales limestone has been variously estimated. Palmer (1938, p. 20) says it ranges from 1700 to at least 5000 feet. For Pinar del Río Province the thickness has been estimated by De Golyer (1918, p. 139) as at least 2500 feet, by Lewis (1932, p. 536-538) as about 2000 feet, and by Dickerson (1937, p. 418) as 1000 to 1500 feet. For Santa Clara Province, Dickerson reports a continuous section of 1300 feet in the low hills northwest of Sagua la Grande. Rutten (1936, p. 29) refrains from estimating the thickness of the aptychi-bearing Viñales limestone of northern Santa Clara Province because of the lack of index horizons and the presence of complicated structures, but his sections show that the limestone is very thick.

LITHOLOGIC FEATURES

The Viñales limestone (Aptychus formation of some authors) was defined by De Golyer (1918, p. 139) as including the hard, blue, massive-appearing limestone forming the central portions of the Sierras Rosario and Organos in Pinar del Río Province. Subsequently, it has been discussed briefly by many writers either in connection with the petroleum possibilities of Cuba, or the age of the formation, or the regional geology of Cuba. These discussions, already sufficiently summarized by Schuchert (1936, p. 514-521), show that the Viñales limestone consists mainly of dark-gray to black, fairly thin-bedded limestone but includes considerable amounts of dark shale and chert. The limestone is compact, generally not visibly crystalline, cut by a fine network of calcite veins, hollowed by many caverns and solution cavities, and weathers bluish or whitish. The chert is black, brownish red, or purple and is generally much fractured. Biologically, the formation is characterized by many aptychi, rare and generally poorly preserved ammonites, abundant Radiolaria, and some plant material. Topographically, it weathers into vertical-walled masses, called mogotes, that rise abruptly to considerable heights above relatively flat-floored valleys underlain by schists and phyllites of the so-called San Cayetano formation. These limestone mogotes, according to Meyerhoff (1938, pp. 280-284), characterize the advanced stages of development of karst topography.

Generalized sections of the Viñales limestone of the various provinces show that the formation remains remarkably uniform lithologically. In the mountains north of Candelaria in Pinar del Río Province, the "top of the Jurassic section is formed by sheer cliffs of hard gray and gray-black, thin-bedded limestones interbedded with thin seams of brownish shale, which weathers white. Some iron is disseminated through the strata and there are occasional lenses of chalcedony" (Brown and O'Connell, 1922, p. 645).

In a company report of October 1933, Butt describes the Viñales limestones of Piner del Río Province as follows:

> "They are dark gray to black in color and occur in beds ranging from one inch to three feet in thickness. Interbedded with them in some places are one to three inch layers of black chert. Locally they are somewhat crystalline but generally they are not, nor do they show evidence of having been subjected to the metamorphosing agencies which have affected the [underlying] San Cayetano formation."

The Viñales limestone of Matanzas Province has been described by W. D. Chawner in a company report as being thin-bedded to massive, friable, locally oölitic and fragmental, and containing much organic shale and some dark chert.

In northern Santa Clara Province the Viñales limestone contains oölites. According to Rutten (1936, p. 10), it

> "is built up largely by limestones and, to a much smaller extent, by cherts, whereas, locally, marls and sandstones or intercalated layers of tuff occur. The limestones are monotonous, dull, grey or greyish-blue, sometimes reddish, dense, and well to finely bedded. Intercalated are layers and lenses of dark dense cherts. In some localities occur grey-brown, fine-grained, sandstones——."

In Camagüey Province the Viñales limestones, according to MacGillavry (1937, p. 7),

> "are rather monotonous and lithologically much the same as those of northern Santa Clara and of Pinar del Río Province. They are buff to greyish blue, fine-grained, compact and generally thin-bedded; they are mostly full of Radiolaria and a certain amount of asphalt seems to be always present. Other members are cherts and radiolarites."

STRATIGRAPHIC BOUNDARIES

The Viñales limestone is overlain with apparent conformity by tuffaceous beds of great thickness containing intercalations of limestone, chert, and shale that become less common toward the top and are probably of early Upper Cretaceous age (Rutter, 1936, p. 7; Vermunt, 1937, p. 15; MacGillavry, 1937, p. 10-12; Thiadens, 1937, p. 11, 12). Judging from formational distribution, Dickerson (1937, p. 419) considers that the Upper Cretaceous beds probably lie unconformably on the Viñales limestone.

The relationship of the Viñales limestone to the underlying group of highly folded, metamorphosed formations, generally referred to as the San Cayetano formation, has been a subject of considerable dispute. Some writers (De Golyer, 1918, p. 140) have insisted that the San Cayetano formation overlies the Viñales limestone, and others (Metcalf, in Lewis, 1932, p. 553; Vermunt, 1937, p. 5-8) have insisted that the Viñales limestone is a facies within the San Cayetano formation, but the opinion of geologists (Lewis, 1932, p. 536; Dickerson and Butt, 1936, p. 116-118) who have done the greatest amount of field work in Cuba, is that the Viñales limestone overlies the San Cayetano formation with angular unconformity. The superposition of the Viñales limestone is shown likewise by the topographic relationship of the limestone to the San Cayetano shales and phyllites (Meyerhoff, 1938, p. 281) and by its fauna being younger than that of the San Cayetano formation.

Evidence for an angular unconformity between these formations consists of differences in degree of metamorphism and folding, and the absence of certain faunas normally present in a complete sequence. Most writers (Brown and O'Connell, 1922, p. 644, 654; Lewis, 1932, p. 534-536; Dickerson and Butt, 1935, p. 116) agree that the San Cayetano schists, phyllites, quartzitic sandstones, and marbles are much more metamorphosed and strongly folded than the Viñales limestone, although Vermunt (1937, p. 5-10) does not believe that the two formations can be separated.

One of the clearest statements on the subject was made by Butt in a company report of October 1933, as follows:

> "I am still of the opinion that the Viñales limestones lie upon the San Cayetano formation with probable angular unconformity. Throughout the Organos Mountains from Guane to San Diego de los Baños the San Cayetano formation is very complexly folded and its members have been considerably metamorphosed. Its sandstone members have become quartzitic and its shale members phyllitic, as a result of the deformational movements to which they have been subjected. The Viñales limestones, in contrast, while considerably faulted, are neither as complexly folded nor do they have the crystalline nature they would undoubtedly show had they the same history as the members of the San Cayetano formation. This difference in complexity of folding and in degree of metamorphism between these two formations cannot be explained, in my opinion, by greater competency on the part of the Viñales formation for there are many sandstone members in the San Cayetano formation which are equally as competent as the thin Viñales limestones."

Likewise bearing on differences in degree of metamorphism is a significant statement by Dickerson (1937, p. 419) as follows:

> "——Jurassic fossils of Oxfordian age have been found in place in limy concretions embedded in phyllites in Pinar del Río Province. At one such locality, La Jaguey Vieja, bitumens were picked from the center of a large ammonite collected from a concretion in place. Upon analysis, the fixed carbon was 79.4 per cent, calculated on an ash-free, water-free basis. Mario Sanchez Roig has published two other analyses of bitumens directly associated with Jurassic fossils. Of these, sample No. 1 gives a carbon ratio of 83.4 per cent, and sample No. 2 [of] 57 5 per cent. These carbon ratios are notably high when compared with those obtained from analyses of bitumens probably yielded by Viñales limestone——. A bitumen collected from a locality near Baracoa gave a carbon ratio of 29 per cent; another from Loma Zambumbia, Santa Clara Province, had a coke residue of 30 per cent. This residue possibly contained 5 per cent ash. The difference between the two sets of carbon ratios and fixed carbon gives a measure of the comparative metamorphism of San Cayetano formation and the Viñales limestone."

Fossil evidence of an angular unconformity between the San Cayetano and Viñales formations, based on the absence of certain faunas that should normally be present if the sequence were complete, was first presented by Dickerson and Butt (1935, p. 116-118). They showed that the previously described Argovian-lower Kimmeridgian fossils, supposedly from the Viñales limestone, actually were derived from the San Cayetano formation but had been mixed in places with float from the overlying Viñales limestone. They claimed that the fossils of the Viñales limestone are of much younger early Lower Cretaceous age, thereby indicating that much of the Upper Jurassic is not represented by sediments in Cuba. The present work substantiates their main contentions, although placing the entire Viñales limestone in the Portlandian stage of the late Jurassic.

ORIGIN

The Viñales limestone was "deposited in quiet waters on the continental shelf" (Schuchert, 1935, p. 514). This is indicated by distinct and generally thin bedding, included plant debris, high bituminous content, and the general absence of structure and texture suggestive of deposition in agitated waters. Deposition was probably in the deeper part of the neritic zone as indicated by the rarity of bottom-living mollusks, by the relatively much greater abundance of ammonites, by the scarcity of sandstones, and perhaps by the presence of layers of Radiolaria-bearing chert. Local shallows subjected to considerable wave action are suggested by the presence of oölites in northern Camagüey and Santa Clara provinces. Deposition must have been fairly rapid considering the great thickness of sediments that accumulated during a minor fraction of the Upper Jurassic.

FOSSILS

LOCALITIES

The fossils examined are from 45 localities of which 11 are in Pinar del Río Province, 29 in Santa Clara Province, and 5 in Camagüey Province. Most of the fossils, including all from Pinar del Río and Santa Clara provinces, were collected by geologists of the Atlantic Refining Company of Cuba; only a few fossils from four localities in Camagüey Province were collected by others. Therefore, it is considered practical and of possible future utility to use the company's laboratory numbers on index maps and fossil lists. The company's collections given to the Geological Survey have received a second set of numbers that will be included in the locality descriptions for purposes of cross reference. The general position of each locality is shown on Figures 1 to 4. Detailed descriptions of localities are given in Table 1:

TABLE 1.—*Fossil localities in the Viñales limestone*

Atlantic Refining Company localities	U. S. G. S. Mesozoic localities	Description of locality, collector, and date
	11285	On farm of Ismael Hernandez, about 1 mile SW. of Minas on railroad between Camagüey and Nuevitas, Camaguey Province. F. L. Wilde, 1922.
60	14455	On north side of main road, between Santa Clara and Camajuani, 5 km. W. of crossing of main road in Camajuani with the Ferrocarril Central, Santa Clara Province. Roy E. Dickerson, 1928.
105	14721	Cantera, 1 km. N. of Zulueta, Santa Clara Province. R. H. Palmer, 1929.
127	14679	Cut, 1 km. from station of Central Carmita, Santa Clara Province. R. H. Palmer, 1929.
142	14722	Loma Yagüey, 3 km. N. of Iguara, about 11 km. S. of Yaguajay, Santa Clara Province. R. H. Palmer, 1929.
178	15351	Vista Hermosa, 8 miles S. of Sagua la Grande, Santa Clara Province. R. H. Palmer, 1929.
180		Loma Manuelita, 6 miles S. of Sagua la Grande, Santa Clara Province. R. H. Palmer, 1929.
184		Hill about 1 km. S. of Sitio Grande and on west side of road, Santa Clara Province. R. H. Palmer, 1929.
205		One-half km. NW. of Rancho Veloz on south side of road, Santa Clara Province. R. H. Palmer, 1929.
211		San Rafael, 4 miles SW. of Cifuentes, Santa Clara Province. R. H. Palmer, 1929.
213		La Trocha, 2 miles SW. of Chinchilla and 5 miles W. of Sagua la Grande, Santa Clara Province. R. H. Palmer, 1929.
215	15352	North slope of Loma Pentón, 6 miles W. of Sagua la Grande, Santa Clara Province. R. H. Palmer, 1929.
217		North slope of Loma Pentón, 6 miles W. of Sagua la Grande, Santa Clara Province. Lower in section than locality 215. R. H. Palmer, 1929.
220, 221		Three and one-half miles NE. of Cifuentes and 1 km. N. of Central Unidad station at Los Angeles, between Cifuentes and Mata, Santa Clara Province. R. H. Palmer, 1929.
226		Six miles E. of Cifuentes and 1.1 miles N. of Central Lutgarda near Mata, Santa Clara Province. R. H. Palmer, 1929.
234	15353	Alacranes, 3 km. W. of Central Santa Teresa which is $4^{1}/_{5}$ miles S. of Sagua la Grande, Santa Clara Province. R. H. Palmer, 1929.
237		Boulders from foot of hill 2½ km. W. of Corralillo, Santa Clara Province. R. H. Palmer, 1929.

TABLE 1.—*Fossil localities in the Viñales limestone*—Continued

Atlantic Refining Company localities	U. S. G. S. Mesozoic localities	Description of locality, collector, and date
246	15357	Colonia Guines, 13 miles NW. of Sagua la Grande on road to San Ramón, Santa Clara Province. R. H. Palmer, 1929.
248		Loma Margarita, 13 miles W. of Sagua la Grande, Santa Clara Province. R. H. Palmer, 1929.
249		One km. E. of Hojalata which is 3.7 km. NE. of Quemado de Guines, Santa Clara Province. R. H. Palmer, 1929.
250	15358	Santa Rita, 5 miles S. of Cifuentes, Santa Clara Province. R. H. Palmer, 1929.
251	15359	Railway cut between O'Reilly and San Francisco, on north end of Loma San Francisco, 16 miles W. and 5 miles N. of Sagua la Grande, Santa Clara Province. R. H. Palmer, 1929.
252	15360	One km. N. of Placetas, Santa Clara Province. R. H. Palmer, 1929.
257	15361	Loma Carmita near Central Carmita which is 12 miles NE of Santa Clara and 2½ miles SE. of Vega Alta, Santa Clara Province. R. H. Palmer, 1929.
259	15362	Small hill on north side of Arroyo Canoa, 1½ miles NW. of Vega Alta, Santa Clara Province. R. H. Palmer, 1929.
268	15363	Northeast slope of Loma Sabanilla, 2½ miles N. of Central Ramona, Santa Clara Province. R. H. Palmer, 1929.
269	15363	West slope of Loma Sabanilla, 2½ miles N. of Central Ramona, Santa Clara Province. Same beds as at locality 268. R. H. Palmer, 1929.
2372		Sierra Camaján, 2.8 kilometers northeast of main road between Minas and Camaguey, about 2 km. W. of Plazolita (known also as Magdalena and Yaguajay), Camaguey Province. Roy E. Dickerson and N. E. Weisbord.
5138	16712	Hill 1200 to 1400 meters N. 50° W. of Puerta del Ancón, Pinar del Río Province. R. E. Dickerson, Emilio Alemán, Antonio Martínez, W. H. Butt, 1933.
5165		Puerta del Ancón, float in field on west side, 100 feet south of highest point on road over pass, Pinar del Río Province. R. E. Dickerson, Emilio Alemán, and Antonio Martínez, 1933.
5211		On El Toro trail, about 400 meters from end, Pinar del Río Province. Roy E. Dickerson and Emilio Alemán, 1934.
5216	16714	Just northwest of La Catalina, a small village about 8 km. NW. of San Diego de los Baños, Pinar del Río Province. Fossils collected not more than 50 feet above base of Viñales limestone, according to Dickerson. Roy E. Dickerson, Emilio Alemán, W. H. Butt, 1934.

TABLE 1.—*Fossil localities in the Viñales limestone*—Continued

Atlantic Refining Company localities	U. S. G. S. Mesozoic localities	Description of locality, collector, and date
5229		On trail 500 meters E. of Hoyo El Rosario and 5 km. N. of Puerta La Muralla which is a place on the mountain front 5 km. N. 70° W. of San Cristóbal, Pinar del Río Province. Roy E. Dickerson and Emilio Alemán, 1934.
5230		In field east of trail on northeast side of Hoyo El Rosario and 700 meters NW. of locality 5229, Pinar del Río Province. Roy E. Dickerson and Emilio Alemán, 1934.
5231		On trail to Hoyo El Rosario on west side of Río San Francisco and 1 km. N. of Puerta La Muralla which is a small place 5 km. N. 70° W. of San Cristóbal, Pinar del Río Province. Roy E. Dickerson and Emilio Alemán, 1934.
5246		Five hundred feet above small tributary to the Río Honda on finca (estate) of Rafael Begoa, Pinar del Río Province. Roy E. Dickerson, 1934.
5256		About 120 meters S. of mountain front and 1000 meters N. 80° E. of house of Mamerto Paz which is 500 meters N. 20° W. of Ginebra, a small store 6 km. N. of Candelaria, Pinar del Río Province. Roy E. Dickerson and Emilio Alemán, 1934.
5264		West side of Canyon of Río Santa Cruz, about 200 feet above river, and 3 km. W. of Ginebra, a small store 6 km. N. of Candelaria, Pinar del Río Province. Roy E. Dickerson and Emilio Alemán, 1934.
5271		Two hundred meters N. 40° E. of chimney of Central Carmita at kilometer 119 on the Tarafa Railroad between Santa Clara and Camajuaní, Santa Clara Province. Roy E. Dickerson, 1934.
5272		Five hundred meters N. of chimney of Central Carmita, Santa Clara Province. Roy E. Dickerson, 1934.
7539		Northwest slope of Loma Sabanilla, 2½ miles N. of Central Ramona, Santa Clara Province. About same locality as 269. D. W. Gravell, John Klecker, and Emilio Alemán, 1941.
	18579	Near shafts sunk for oil on southern slope of a hill on the Regla Finca, 3 to 4 km. W. of Minas on the Camaguey-Nuevitas branch of the Cuban railroad, Camagüey Province. Fossils from conglomerate zone in limestone. Thomas P. Thayer, 1941.
	18580	Five hundred feet away from locality 18579. Thomas P. Thayer, 1941.
	18581	Just inside boundary of Regla Finca on north slope of low hill about three-fourths of a mile E. of locality 18579. Thomas P. Thayer, 1941.
3083	18653	Baños de Soroa, about 5 miles NW. of Candelaria, Pinar del Río Province. Probably collected by N. E. Weisbord.

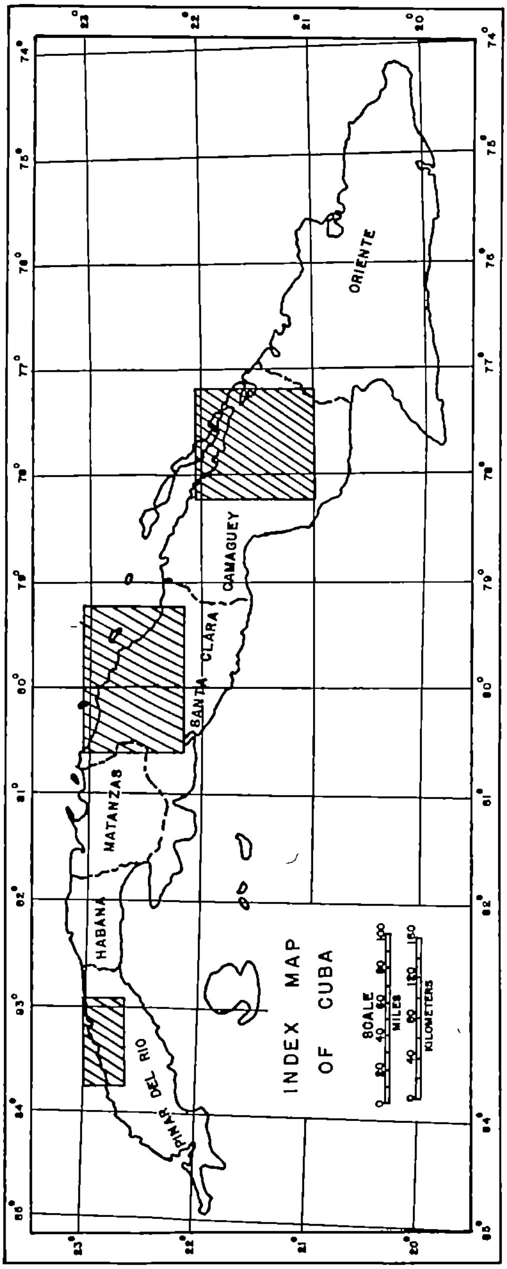

FIGURE 1.—*Index map of Cuba*

Hatched areas show positions of fossil localities indicated in Figures 2 to 4.

TABLE 2.—*Geographic distribution of the late Jurassic fossils of the Viñales limestone of Cuba*

Cuban Provinces	Santa Clara																													Pinar del Río										Camaguey				
U. S. G. S. Mesozoic localities	14455	14721	14679	14722	15351						15352				15353		15357			15358	15359	15360	15361	15362	15363	18653				16712			16714							11285		18579	18580	18581
Atlantic Refining Company of Cuba localities	60	105	127	142	178	180	184	205	211	213	215	217	220 & 221	226	234	237	246	248	249	250	251	252	257	259	268 & 269	3083	5271	5272	7539	5138	5165	5211	5216	5229	5230	5231	5246	5256	5264		2372			
Phylloceras pinarense Imlay																																	X											
P. sp.																X																												
Metahaploceras cf. *M. mazapilense* (Burckhardt)																																	X											
M. sp.																													X															
Pseudolissoceras cf. *P. zitteli* (Burckhardt)												X													X				X				X	X		X								
P. sp.										X																															?		?	?
Hildoglochiceras cf. *H. grossicostatum* Imlay																														X														
H. cf. *H. alamense* Imlay																														X														
H. sp.																									?				X															
Simoceras sp. juv. cf. *S. volanense* (Oppel)																									X																			
Virgatosimoceras? sp.																																	X											
Aspidoceras spp.																									X								X											
Physodoceras sp.																																	X											
Virgatosphinctes cristobalensis Imlay																																		X										
V. cf. *V. cristobalensis* Imlay																																					X							
V. aff. *V. rotundidoma* Uhlig																																	X	X		X								
Corongoceras filicostatum Imlay																		X							X				X					X										
C. cf. *filicostatum* Imlay																																		X										
Dickersonia sabanillensis Imlay									?		X	X													X				X															?
D. ramonensis Imlay		?																							X				X						?	?								
Micracanthoceras sp. juv.																									X				X															
Durangites cf. *D. acanthicus* Burckhardt																									X																			
D. vulgaris Burckhardt																																												X
Lytohoplites caribbeanus Imlay																																		X										
Parodontoceras butti Imlay																																	X	X		X								
P. antilleanum Imlay																			X														X	X		?								
P. sp.																								X							X				X			X	X					X
Berriasella? sp.																									X								X											
Spiticeras? sp.																											X																	
Leptoceras? hondense Imlay																							X					X					X				X							
L. catalinense Imlay																							X				?	X					X											
L. sp.						X		X												X																								
Hamulina? rosariensis Imlay										?																							X	X		X							?	
Ptychoceras? sp.							X																																					
Lamellaptychus rectecostatus (Peters) emend. Trauth																										X																		
L. angulocostatus (Peters)				X							X					X				X			X			X	X																	
L. angulocostatus var. *cristobalensis* (O'Connell)				X																						X							?											
L. seranonis (Coquand)				X	X																X	X					?					X								X		X		
L. excavatus Trauth	X																					X																						
L. sp.					X	X		X	X				X		X		X		X					X	X						X					X	X							X
Hadrocheilus sp.														X																														
Pleurotomaria? sp.																X																												
Brachiopod																									X																			
Echinoid spine																									X																			
Coral																X																												
Sponge														X																														

STATE OF PRESERVATION

The fossils of the Viñales limestone consist mainly of poorly preserved, distorted, internal molds retaining some shell material. A few silicified fossils are fairly well preserved, although having porous surfaces that do not permit magnification. Suture lines generally cannot be traced. Specifically determinable fossils available for the present study include about 400 normal ammonites, 150 uncoiled ammonites, and 65 aptychi.

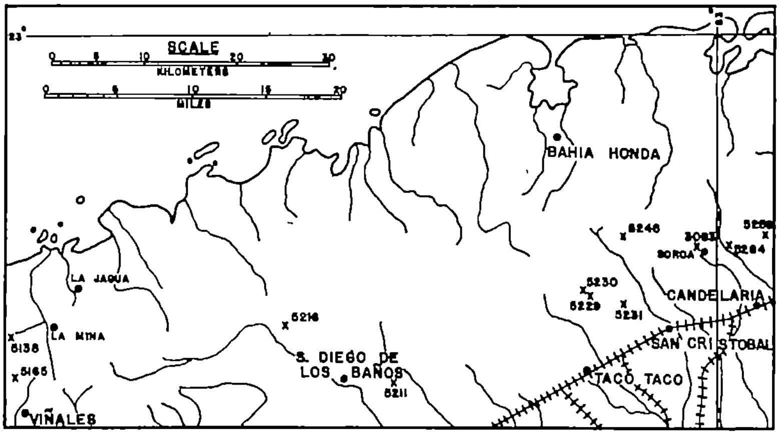

FIGURE 2.—*Map of fossil localities in Pinar del Río Province*
All numbers represent localities of the Atlantic Refining Company of Cuba.

ANALYSIS OF FAUNA

The Viñales limestone fauna discussed herein (Table 2) includes 23 species of normally coiled ammonites, 4 of uncoiled ammonites, 7 of aptychi, and 1 cephalopod beak. The gastropods, pelecypods, brachiopods, echinoids, corals, and sponges are represented by a few fragmentary specimens not worth description. The cephalopods are distributed among 22 genera of which one is new. Of the ammonites, 11 species are new, 5 are identical or closely comparable with Mexican species, 1 is compared with an Argentinian species, 2 are compared with Alpine-Mediterranean species, and 8 are not specifically determinable. Most of the aptychi are identical with species from southern Europe. The affinities of the normal ammonites are predominantly with the late Jurassic ammonites of Mexico and Argentina and to a lesser extent with those of the Alpine-Mediterranean and Indian provinces. The affinities of the uncoiled ammonites and aptychi are predominantly with the early Lower Cretaceous faunas of the Alpine-Mediterranean province and slightly with any known late Jurassic faunas.

The Phylloceratidae are represented by one species of *Phylloceras;* the Oppelidae by two species of *Metahaploceras;* the Haploceratidae by two species of *Hildoglochiceras* and one of *Pseudolissoceras;* the Aspidoceratidae by a few immature or fragmentary specimens of *Simoceras,* *Virgatosimoceras?*, *Aspidoceras*, and *Physodoceras;* and the Perisphinctidae by two species of *Virgatosphinctes.* The Berriasellidae are well represented by seven genera and nine species. *Corongoceras* is well known from Argentina, Algeria, and southern Europe but has not been recorded previously from North America. *Dickersonia,* n. gen., includes two species whose inner whorls are similar to *Corongoceras* but whose outer whorls develop simple *Berriasella*-like ribbing and appear to be most similar to forms from the latest Jurassic of India. *Micracanthoceras* is represented by only a few immature forms. *Durangites* includes two species closely related to or identical with Mexican species. The genus has not been recorded previously outside Mexico. *Lytohoplites* is referred to the Berriasellidae rather than the Neocomitidae, because of its rounded whorl section and the persistence of its ventral groove. Its occurrence in Cuba is of unusual interest, as it has been recorded previously only from Argentina, where it is confined to a narrow zone at the

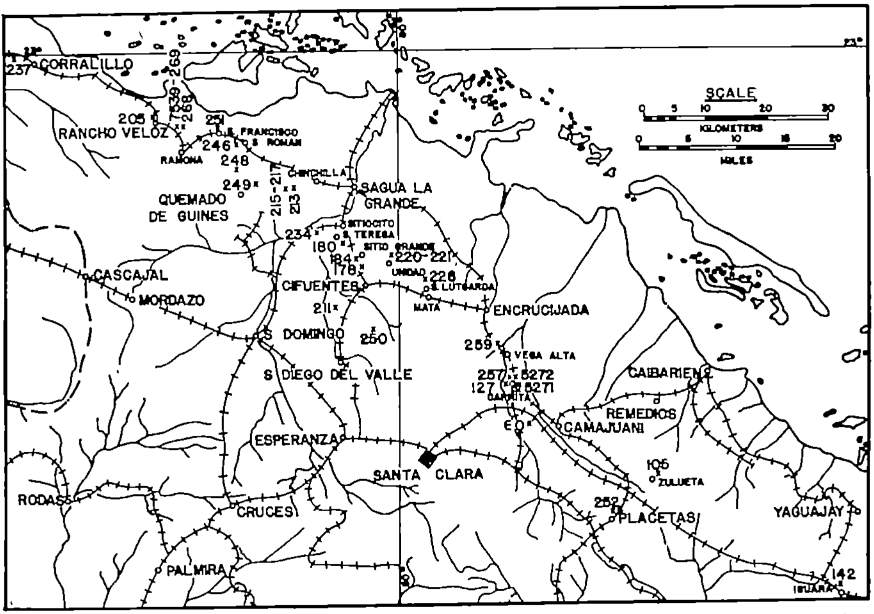

FIGURE 3—*Map of fossil localities in Santa Clara Province*
All numbers represent localities of the Atlantic Refining Company of Cuba

Portlandian-Tithonian boundary. *Parodontoceras* includes two species characterized by unbranched ribs in the adult and comparable with species from the Tithonian of Mexico and the Berriasian of France. *Berriasella* is doubtfully represented by one species. The Olcostephanidae are represented by one specimen doubtfully referred to *Spiticeras*.

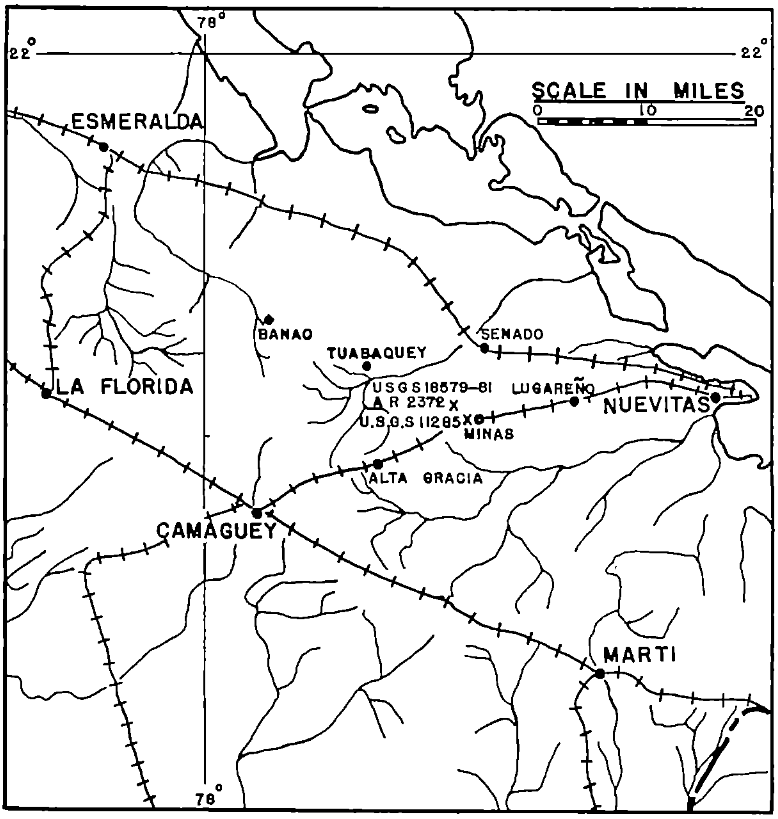

FIGURE 4.—*Map of fossil localities in Camaguey Province*

Only number 2372 represents a locality of the Atlantic Refining Company of Cuba. The remaining numbers represent localities for the U. S Geological Survey.

The family relationships of the uncoiled ammonites are highly uncertain, but provisionally the Cuban species referred to *Leptoceras?* may be assigned to the Ancyloceratidae, and the species referred to *Hamulina?* and *Ptychoceras?* may be assigned to the Lytoceratidae.

All the aptychi in the Cuban collections belong to the form genus *Lamellaptychus* Trauth. The single cephalopod beak is identified with *Hadrocheilus* Till.

TABLE 3.—*Similar Jurassic species in Cuba, Mexico, Argentina, and Eurasia*

Species from the Viñales limestone of Cuba	Similar species	Occurrence of similar species	Age of similar species
Phylloceras pinarense Imlay	*P. serum* (Oppel)	Southern Eruope	Late Jurassic
Metahaploceras cf. *M. mazapilense* (Burckhardt)	*M. mazapilense* (Burckhardt)	Zacatecas, Mexico	Portlandian
Hildoglochiceras cf. *H. grossicostatum* (Imlay)	*H. grossicostatum* Imlay	Coahula, Mexico	Upper Portlandian
H. cf. *H. alamense* Imlay	*H. alamense* Imlay	Coahuila, Mexico	Tithonian
Pseudolissoceras cf. *P. zitteli* (Burckhardt)	*P. zitteli* (Burckhardt)	Argentina	Lower Portlandian
Simoceras cf. *S. volanense* (Oppel)	*S. volanense* (Oppel)	Southern Europe	Portlandian
Virgatosphinctes cristobalensis Imlay	*V. frequens* (Oppel)	India	Portlandian
V. aff. *V. rotundidoma* Uhlig	*V. rotundidoma* Uhlig	India	Portlandian
Corongoceras filicostatum Imlay	*C. alternans* (Gerth)	Argentina	Lower Tithonian
Durangites cf. *D. acanthicus* Burckhardt	*D. acanthicus* Burckhardt	Durango, Mexico	Upper Portlandian
D. vulgaris Burckhardt	*D. vulgaris* Burckhardt	Durango, Mexico	Upper Portlandian
Lytohoplites caribbeanus Imlay	*L. vetustoides* (Burckhardt)	Argentina	Upper Portlandian
Parodontoceras butti Imlay	*P. simplicicostatum* (Mazenot)	France	Berriasian
P. antilleanum Imlay	*P.* cf. *P. calistoides* (Behrendsen)	Zacatecas, Mexico	Tithonian
Leptoceras? hondense Imlay	*L.? gracile* (Oppel)	Southern Europe	Portlandian
L ? catalinense Imlay	*L.* sp. ind. Mazenot	France	Berriasian

CORRELATION

The Viñales limestone fauna is correlated with the upper part of the Portlandian stage of the late Jurassic on the basis of the known ranges of its normally coiled ammonite genera and the similarity, or identity, of some of its species with upper Portlandian species of Mexico and Argentina. The ranges of some of the genera are as follows:

Metahaploceras, Kimmeridgian and Portlandian.
Hildoglochiceras, upper Kimmeridgian?, Portlandian, and Tithonian.
Pseudolissoceras, Portlandian.
Simoceras, Portlandian and Tithonian.
Aspidoceras, Kimmeridgian to Tithonian.
Physodoceras, Kimmeridgian and Portlandian.
Virgatosphinctes, upper Kimmeridgian to Tithonian.
Corongoceras, upper Portlandian and lower Tithonian.
Micracanthoceras, upper Portlandian and Tithonian.
Durangites, upper Portlandian.
Lytohoplites, upper Portlandian and lower Tithonian.
Parodontoceras, upper Portlandian to Berriasian.

The Portlandian age of the above assemblage is indisputable, although genera, such as *Pseudolissoceras*, *Corongoceras*, and *Lytohoplites*, known from only a few species, may have longer ranges than indicated. *Metahaploceras* is common in the Kimmeridgian and rare in Portlandian. *Hildoglochiceras* is common in the Portlandian, rare in the Tithonian, and questionably present in the Kimmeridgian. *Pseudolissoceras* is associated with *Subplanites* and *Virgatosphinctes* in the lower Portlandian of Argentina and Mexico, but in Europe it appears to range through the Portlandian. *Simoceras* is apparently most common in the Tithonian. *Aspidoceras* is a characteristic late Upper Jurassic genus. *Physodoceras* is rare above the Kimmeridgian. *Virgatosphinctes* is rare below the Portlandian. *Corongoceras* in Argentina ranges from just above the beds with *Pseudolissoceras* to the lowest beds with *Substeueroceras*. *Durangites* has been found at many places in Mexico, either associated with *Kossmatia*, or just below beds containing *Substeueroceras*. Its presence in Cuba is considered very strong evidence of the upper Portlandian age of the beds in which it occurs. *Lytohoplites* has been reported previously only from Argentina, where it occurs in the lowest bed containing *Substeueroceras* and *Protacanthodiscus* as well as in the underlying bed. *Parodontoceras* in Mexico is not known below the Tithonian but in Argentina is reported also in the upper Portlandian just above beds containing *Pseudolissoceras*.

Further evidence of a Portlandian age is furnished by comparisons of the Cuban species with the most similar species of Mexico, Argentina, and Eurasia, as shown in Table 3.

An upper rather than lower Portlandian age for the Viñales limestone is indicated by the ranges of the ammonite genera and of the similar species. *Corongoceras, Micracanthoceras, Durangites, Lytohoplites,* and *Parodontoceras* are not known below the upper Portlandian. *Metahaploceras, Physodoceras,* and *Durangites* are not known above the Portlandian. *Corongoceras* and *Lytohoplites* are not known above the basal beds of Tithonian age in Argentina. Most of the similar species are upper Portlandian or younger. The only suggestion of a lower Portlandian age for the Viñales limestone is furnished by specimens of *Pseudolissoceras* comparable with species from Argentina and Mexico, where the genus is known only from the lower Portlandian. However, in Europe *Pseudolissoceras* is not so narrowly restricted, and in Cuba specimens showing the simple sutures characteristic of the genus were obtained at localities 5216 and 7539 in association with other genera not known below the upper Portlandian.

The fossil assemblage at locality 5216, near La Catalina in Pinar del Río Province, is of particular stratigraphic importance, as it was obtained, according to Roy E. Dickerson (personal communication), only a few feet above the base of the Viñales limestone. Most of the species at locality 5216 occur likewise in the same region at localities 5231 and 5229 which Dickerson considers are some hundreds of feet higher stratigraphically. Furthermore, all other fossil localities of the Viñales limestone of Cuba show a community of species with the localities mentioned above. Most of them are definitely upper Portlandian, and none indicates lower Portlandian. Therefore, it may be doubted whether any part of the Viñales limestone is lower Portlandian.

Nothing in the fauna of the Viñales limestone suggests the Kimmeridgian stage. Even the long-ranging species of aptychi are entirely unlike the aptychi in the *Mazapilites* and *Idoceras* beds of the Kimmeridgian of Mexico, judging by about 25 specimens in the collections of the University of Michigan. Likewise, nothing in the fauna suggests the Tithonian stage of latest Jurassic, which is represented in Mexico and Argentina by beds containing many species of *Substeueroceras.* Considering the reported great thickness of the Viñales limestone, its restriction to the upper Portlandian is surprising, although in the Placer de Guadalupe District of Chihuahua, Mexico, the Portlandian is represented by a similar great thickness. However, the presence of late Upper Jurassic rocks in Cuba is not surprising as rocks of that age are known from Trinidad (Spath, 1939, p. 187-189) and from the Gulf region of Mexico (Burckhardt, 1930, p. 94-96, 266; Muir, 1936, p. 11-17; Kellum, 1937, p. 39, 40, 69-71, 86-91, Pl. 9, figs. 1-6, 12).

Previous correlations (Schuchert, 1935, p. 514-516; Dickerson and Butt, 1935, p. 116-118; Trauth, 1936, p. 66-68; Rutten, 1936, p. 10; Vermunt, 1937, p. 5, 8, 12; MacGillavry, 1937, p. 7, 8) with the Lower Cretaceous were based mainly on the identity of the aptychi and the similarity of the uncoiled ammonites to species in the early Lower Cretaceous deposits of the Alpine-Mediterranean region. Among the six identical species and varieties of aptychi, three have not been reported previously from beds older than the Lower Cretaceous. The richness in form and number of uncoiled ammonites in the Viñales limestone of Cuba contrasts markedly with the scarcity of uncoiled ammonites in Upper Jurassic deposits of other parts of the world but compares favorably with their abundance in the Lower Cretaceous. Association of the aptychi and uncoiled ammonites with normally coiled ammonites in the same beds of the Viñales limestone permits no doubt as to their Portlandian age and suggests that aptychi must be used with caution in detailed stratigraphic work. This unusual association in Cuba probably explains the perplexities of some geologists (O'Connell, 1921, p. 1; Lewis, 1932, p. 943, 944; Vermunt, 1937, p. 5-12) who found limestones containing supposed Cretaceous aptychi intercalated with shales containing Upper Jurassic ammonites.

Further confusion and controversy concerning the age of the Viñales limestone has resulted from the assumption by collectors that the Argovian-lower Kimmeridgian ammonites described by Sanchez Roig (1920) and O'Connell (1920) were obtained from the Viñales limestone or from its float. This assumption was based on the fact that most of the fossils were found on the slopes bordering the vertical-walled hills of Viñales limestone. It was corrected by Dickerson and Butt (1935, p. 116-118) who, on the basis of considerable field work, determined that the ammonites actually occur in the phyllites of the San Cayetano formation but have been mixed locally with float from the overlying Viñales limestone. At many places, they collected older Jurassic (Argovian-lower Kimmeridgian) ammonites from concretions in the San Cayetano formation. Nowhere did they observe similar concretions in the Viñales limestone. Their observations and contentions are substantiated by the present study which shows that the Viñales limestone is not older than the Portlandian, is probably not older at its base than the upper Portlandian, and is decidedly younger than the Argovian-lower Kimmeridgian fossils previously described.

The ammonites of the San Cayetano formation, described in part by Sanchez Roig and O'Connell, have been discussed by Burckhardt (1930, p. 61, 62), Spath (1931, p. 400, 592, 593), and Arkell (1939, p. LXIV) and have been studied considerably in both Europe and America. The fossils of Sanchez Roig have been examined and commented on independently

TABLE 4.—*Upper Jurassic faunal divisions of Mexico, Argentina, and Cuba*

European stages		Mexico		Argentina		Cuba
Tithonian		*Substeueroceras* and *Proniceras*	*Micracanthoceras* *Hildoglochiceras*	Substeueroceras	*Micracanthoceras*, *Parodontoceras*	Not known
Portlandian	Upper	*Kossmatia* and *Durangites*		*Lytohoplites*, *Kossmatia*, and *Corongoceras*		*Durangites*, *Corongoceras* *Lytohoplites*, *Hildoglochiceras*, *Pseudolissoceras*, *Micracanthoceras*
	Lower	*Subplanites* and *Pseudolissoceras*	*Aulacosphinctoides*	*Pseudolissoceras* and *Subplanites*		Probably absent
Kimmeridgian	Bononian	*Mazapilites*		Not known or doubtfully represented		Not known
	Havrian	*Waagenia*				
		Glochiceras fialar and *Idoceras* group of *durangense*				
	Sequanian	*Idoceras* group of *balderus*				
		Sutneria group of *platynota*				*Ataxioceras*
Oxfordian	Argovian	*Dichotomosphinctes*				*Dichotomosphinctes*, *Biplices*, and *Kranaosphinctes*
	Divesian	*Indosphinctes*?				Not known
Callovian		*Peltoceras*		*Peltoceras* and *Neuqueniceras*		

by Burckhardt and Spath, who seem to agree as to their age. Concerning these fossils Burckhardt says (translation):

> "An examination of this material shows at once that many determinations by M. Sanchez Roig should be rectified. Thus *Perisphinctes lagunitasensis* (pl. I, fig. 1, 1c) does not belong to that species, but to the group of *P. promiscuus* Buk.; *P. durangensis* (I, 2) does not belong to that species, *P. durangensis* (II, 4) belongs to the group of *P. polyplocus* Rein., *P.* aff *wartaeformis* (III, 6; VI, 1) belongs to the group of *P. lictor* Font.; *P. delgadoi* (IV, 2) belongs to the group of *P. frequens* Oppel; *P.* aff. *elisabethae* (IV, 3) is near to *P. frequens* Oppel, *P.* cf. *biplex* (V, 3-4) is closely related to *P. triplicatus albus* Qust. The *Simbirskites* (IX, 1-4) are *Perisphinctes,* in part (2, 4) of the group of *P. virgulatus* Qust.; *Kossmatia* (X, 1-1a) and *Idoceras* (X, 6-7; XI, 1-5) do not belong to these genera but are true *Perisphinctes; Idoceras aguilerae* (X, 6) belongs to the group of *Perisphinctes polygyratus* Rein., *Berriasella* aff. *oppelli* (XII, 1, 1a-b) is a *Perisphinctes* equally of the group of *P. polygyratus, Haploceras fialar* (XIV, 3-5) belongs to the group of *Oppelia lingulata* Qust., *Nebrodites* (XIV, 1-2) and the *Aucellas* are doubtful; finally *Perisphinctes virgulatus* (VIII, 5-5a) is related to *P. frequens* Oppel.
>
> "After these rectifications, the existence of the middle and upper Kimmeridgian and of the Portlandian at Viñales has not been proven as yet. On the other hand the ammonites examined by me indicate with certainty the presence of the zone of *Peltoceras bimammatum (Perisphinctes* cf. *lucingensis* Choffat, *P.* aff. *biplex rotundus* Qust., *P.* aff. *virgulatus* Qust., *P.* aff. *triplicatus albus* Qust. in Engel) and of the zone of *Persphinctes polyplocus (P.* aff. *capillaceous* Font. [pl. IX, 4, 1 c], *P.* gr. *lictor* Font., *P.* aff. *frequens* Oppel, *P.* aff. *polyplocus* Rein.)."

Some of the names used by Sanchez Roig have been corrected by Spath (1931, p. 400, 592, 593) as follows:

CORRECTION BY SPATH	USAGE BY SANCHEZ ROIG
Vinalesphinctes roigi Spath	*Aspidoceras* sp., p. 30, Pl. 12, fig 2.
Vinalesphinctes niger Spath	*Perisphinctes* cf. *colubrinus* Reinecke, p. 19, Pl. 4, fig. 1.
Prososphinctes n. sp.	*Nebrodites* aff. *agrigentinus* E. Favre, p. 31, Pl. 14, figs. 2, 2a.
Prososphinctes n. sp.	*Kossmatia zacatecana* Burckhardt, p. 44, Pl. 10, figs. 1, 1a.
Prososphinctes subconsociatus Spath.	*Perisphinctes?*, p. 23, Pl. 7, figs. 3, 3a.
Biplices? sp.	*Perisphinctes lagunitasensis* Burckhardt, p. 12, Pl. 1, figs. 1, 1a.
Euaspidoceras vinalense Spath	*Aspidoceras* aff. *laevigatum* Burckhardt, p. 29, Pl. 13, fig. 2.
Euaspidoceras o'connelli (Sanchez Roig)	*Aspidoceras o'connelli* Sanchez Roig, p. 30, Pl. 13, figs. 1, 1a.

The groups of perisphinctids mentioned by Burckhardt in the above quotation have been discussed in considerable detail by Spath (1931, p. 397-403, 443-446) and by Arkell (1936, p. XXXIII-XLVI; 1937, p. XLVII-LIV; 1939, p. LV-LXIV; 1940, p. LXV) and have been placed in various genera and subgenera. The group of *Perisphinctes promiscuus* Bukowski has been placed in *Biplices* by Spath and in *Kranaosphinctes* by Arkell. The group of *P. polygyratus* (Reinecke) has been placed in *Planites* by Spath and in *Biplices* by Arkell, who rejects *Planites* as an invalid name. The groups of *P. polyplocus* (Reinecke) and *P. lictor* (Fontannes) belong to *Ataxioceras*. The group of *P. virgulatus* (Quenstedt) belongs to *Discosphinctes*. The forms that Burckhardt referred to the group of *P. frequens* Oppel (*i. e., Virgatosphinctes*) are certainly

TABLE 5.—*Correlation of the Upper Jurassic rocks of Cuba*

European stages		English formations (after Arkell, 1933)	Northern Mexico (After Imlay, 1939)	Southern Mexico (Taken from Burckhardt, 1930)	Arkansas, Louisiana, and eastern Texas (After Imlay, 1941)	Cuba
Tithonian		Purbeck beds	La Casita sandstones, shales, marls, coal, and gypsum	Limestone, marls, and shales in Vera Cruz, Puebla, Chiapas, and Oaxaca	Cotton Valley shales and sandstones	Not known ?
Portlandian		Portland beds				Viñales limestone
Kimmeridgian	Bononian	Kimmeridge clay	(La Caja marls and limestone offshore)		Basal conglomerate	Probably absent
	Havrian		Locally conglomerates or thick masses of gypsum at base	Locally basal conglomerates and sandstones		
	Sequanian				Buckner red beds and anhydrite ?	?
Oxfordian	Argovian	Corallian beds	? La Gloria sandstones and limestones		? Smackover limestone	San Cayetano formation
	Divesian	Oxford clay	(Zuloaga limestone offshore)	? Salt and red beds in Chiapas, Vera Cruz and Oaxaca ?	? Eagle Mills red beds and salt ?	? Not known
Callovian		Kellaways beds	Lower limit not known	Shales, marls, and limestones in Oaxaca and Guerrero		

incorrectly referred and probably belong to *Dichotomosphinctes* or *Discosphinctes*. The forms described by O'Connell (1920, p. 648-680, Pls. 34-36 in part) as *Perisphinctes cubanensis, P. delatorii,* and *P. plicatiloides* should probably be referred to *Dichotomosphinctes* rather than *Biplices* as indicated by the forward curvature of their secondary ribs.

All these perisphinctid genera occur in the upper Oxfordian (Argovian) stage, and most of them do not range above it. However, *Ataxioceras* and *Discosphinctes* attain their greatest development in the lower Kimmeridgian. Whether Kimmeridgian is represented in the fauna of the San Cayetano formation will have to be determined by detailed study, but certainly none of the described species suggests an age as young as the zone of *Idoceras balderus* of late lower Kimmeridgian.

Comparisons of the Upper Jurassic faunas of Cuba with equivalent faunas of Mexico and Argentina (Table 4) indicate that the Cuban section is much less complete than sections in other regions bordering the Gulf of Mexico (Table 5) and that the unconformity between the San Cayetano and Viñales formations involves parts of the Kimmeridgian and Portlandian stages. Possibly future fossil discoveries will show that the Cuban Jurassic includes beds of other ages than now recognized, but present evidence shows that the Kimmeridgian was a time of uplift in Cuba, as in many parts of North America.

Thus the Nevadian orogeny of the west coast, according to Reed (1941, p. 106) and Taliaferro (1941, p. 125, 134, 151), occurred after the deposition of the Mariposa slate containing Kimmeridgian fossils and before the deposition of the Knoxville formation containing Portlandian fossils. The Mariposa slate includes Kimmeridgian species of *Amoeboceras,* which, according to Spath (1933, p. 864; 1935, p. 72), ranges from the middle part of the upper Oxfordian (Argovian) to the lower part of the middle Kimmeridgian. The presence of this ammonite indicates that the Nevadian orogeny did not reach its greatest intensity until late Kimmeridgian time.

Likewise, in the western interior the Kimmeridgian was a time of uplift marking the beginning of deposition of terrestrial deposits of the Morrison formation (Baker, Dane, and Reeside, 1936, p. 9, 58-63). It is extremely doubtful whether any of the underlying marine deposits are as young as Kimmeridgian.

In Mexico marked changes in the characteristics of the sediments occurred in late Oxfordian or early Kimmeridgian time. The light-gray, thick-bedded limestone and yellowish sandstone of Oxfordian age were overlain by large amounts of dark shale, marl, sandstone, and locally gypsiferous and carbonaceous deposits that probably reflected topographic and climatic changes on the land masses surrounding the Mexican sea

(Imlay, 1940b, p 393-396). An unconformity in the marginal areas of the Mexican geosyncline during earliest Kimmeridgian time is suggested by thick masses of gypsum in the Sierra Madre Oriental, by conglomerates in the southern Coahuila, and by the apparent absence in northern Mexico of beds containing *Sutneria* of the group of *S. platynota* Reinecke.

In the Gulf region of the United States the unconformity at the base of the Cotton Valley formation was probably developed during late Oxfordian or early Kimmeridgian time, judging by the fossils obtained from the underlying Smackover limestone. The large quantities of gravel throughout the lower part of the Cotton Valley formation imply much higher bordering land masses than during the deposition of the Smackover limestone (Imlay, 1940a, p. 16-27; 1941, p. 256-261). Furthermore, the coarsening of these formations toward the north and east indicates the source of the sediments (Imlay, 1940a, p. 15, 21, 27) and suggests that the region of the present Southern Appalachian Mountains was a highland during Upper Jurassic time. Possibly the Palisade disturbance, which formed block mountains in eastern North America from Nova Scotia to the Carolinas, occurred during the Upper Jurassic. At present, it is dated as being younger than the Newark group of Upper Triassic age and older than the Potomac group of early Lower Cretaceous age. However, the evidence of: (1) a pronounced unconformity in the middle of the Upper Jurassic sequence of Cuba and of the Gulf region of the United States and (2) highlands east and north of the Gulf Coast Jurassic sea suggest a late rather than early Jurassic age for the disturbance. There was ample time during latest Upper Jurassic and Lower Cretaceous for peneplanation of the highlands before marine transgression across the Atlantic Coast in the early Upper Cretaceous.

ECONOMIC SIGNIFICANCE

Study of the fossils of the Viñales limestone of Cuba has led to conclusions of potential economic significance: (1) The oil-bearing Viñales limestone is of late Upper Jurassic age; (2) part of the underlying metamorphosed San Cayetano formation is of early Upper Jurassic age; (3) an unconformity is present between these formations and is of Kimmeridgian-lower Portlandian age, corresponding roughly with the unconformity developed during the Nevadian orogeny of the Pacific Coast region; (4) no undoubted Lower Cretaceous has yet been discovered in the West Indies; and (5) the Jurassic history of Cuba is similar in major features to that of other regions bordering the Gulf of Mexico. Application of these conclusions toward forming a clearer picture of Jurassic and Lower Cretaceous events should aid in the search for mineral deposits by showing their time of origin and possible distribution. Recognition of an important unconformity in the middle of the petroliferous Upper Jurassic sequence should be of particular value in explorations for petroleum in the deeply buried rocks of the lands bordering the Gulf of Mexico. The unmetamorphosed Viñales limestone is a potential source of commercial petroleum.

In Cuba these conclusions will be useful in dating the intrusions of metalliferous serpentines and granitoid rocks and thereby defining the areas in which ore deposits may be found. Knowledge of the age of igneous intrusions in structurally complex areas will help the petroleum geologist to determine the kind and throw of faults and the probable depth at which the petroliferous Viñales limestones may be encountered. The position of the Viñales limestone is of prime importance in considering the possible presence of oil in reservoir rocks of Upper Cretaceous or Tertiary age. Confirmation of the occurrence of unconformities both above and below the Viñales limestone may likewise have a bearing on oil exploration.

The Upper Jurassic rocks of the Gulf region of the United States have furnished large quantities of petroleum only in southern Arkansas and northern Louisiana but are possible sources of petroleum in an arcuate belt extending from Del Río, Texas, northeastward to Arkansas and from there southeastward to southwestern Alabama. The probability of finding suitable reservoir beds throughout this area is increased by the recognition of pronounced orogenic movements during the Upper Jurassic in the lands bordering the Gulf region. Judging from the Upper Jurassic section of Arkansas and Louisiana, a considerable depression of the Gulf of Mexico occurred during the same time as uplift in the surrounding land masses and permitted the accumulation of a great thickness of sediments. Although the Upper Jurassic orogeny may have been greatest in Cuba, its effects from the viewpoint of petroleum accumulation may be greater elsewhere.

The Upper Jurassic rocks of the Gulf region of Mexico are petroliferous but have not been adequately tested for oil production. The sequence of formations is remarkably similar to that of Arkansas and Louisiana, and the geologic history must likewise have been very similar. Of particular interest is the marked change in conditions of sedimentation at the end of Oxfordian time, or the beginning of Kimmeridgian time, which resulted in thick-bedded limestones being overlain by coarse clastic sediments that include gypsum and coal in their basal part. The change appears to coincide with the development of the disconformity between the Smackover limestone and the Cotton Valley formation of Arkansas and to be distantly related to the middle Upper Jurassic orogeny of Cuba. Detailed studies of the Jurassic rocks exposed in the mountains of eastern

Mexico would be very valuable to geologists interested in petroleum possibilities of the deeply buried Jurassic rocks of the Gulf region.

SYSTEMATIC DESCRIPTIONS

Genus *Phylloceras* Suess, 1865

Phylloceras pinarense Imlay, n. sp.

(Plate 1, figures 1-9)

This species is represented in the collections by about 100 specimens and is the most common of the normally coiled ammonites. Form discoidal, compressed. Whorls elongate-ovate, higher than wide, becoming relatively higher during growth, thickest on lower fourth of flanks, embracing preceding whorls almost completely. Flanks broad, gently convex. Venter narrowly rounded, becoming narrower during growth. Umbilicus narrow, becoming wider during growth, wall steeply inclined, rounding rather abruptly into flanks.

The inner whorls (Pl. 1, figs. 1-4) are ornamented with fine, closely-spaced riblets that cross the flanks nearly radially and are strongest on the venter. These riblets become fainter anteriorly and are not visible on the lower part of the flanks at diameters greater than about 25 mm. They persist on the venter and on the upper part of the flanks at greater diameters but gradually change to striae that tend to be bundled and are rather strongly inclined forward. No radial swelling or furrows present at any growth stage.

Accurate dimensions are difficult to obtain, because many of the specimens have been compressed, and most of them have been much weathered on one side The specimen shown on Figures 6-9 of Plate 1 appears to be nearly undeformed. At a diameter of 78 mm. its whorl height is 44 mm., its whorl thickness is 27 mm., and its umbilical width on the left side is 13 mm.

This species is unlike any described species from North or South America. It shows considerable resemblance to *Phylloceras serum* (Oppel) (*in* Zittel, 1868, p. 66, Pl. 7, figs 5a-c, 6a-c) from the late Jurassic of southern Europe, but the adult has a wider umbilicus, a more narrowly rounded venter, and weaker ribbing. The wider umbilicus is probably due, in part, to the absence of shell material. *Phylloceras tethys D'Orbigny* (1841, p. 174, Pl. 53, figs. 7-9, Pl. 41, figs. 3, 4; Pictet and Loriol, 1858, p. 17, Pl. 3, figs. 1a, b) from the Neocomian of southern Europe is likewise very similar but apparently has stronger ribbing and a more evenly rounded venter on the adult whorls.

TYPES: Holotype, U. S. National Museum 103378; 3 figured paratypes, U. S. National Museum 103379a-c; 13 nonfigured paratypes, U. S. National Museum 103380a-m.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba locality 5216.

Genus *Metahaploceras* Spath, 1925

Metahaploceras cf. *M. mazapilense* (Burckhardt)

(Plate 2, figures 11-13)

One internal mold is tentatively referred to *Metahaploceras*. Whorls elongate-ovate in section, much higher than wide, thickest in the lower third of the flanks, embracing preceding whorls almost completely; flanks broad and nearly flat; venter narrowly rounded. Umbilicus very narrow, wall low and vertical.

The ribbing is falciform but poorly preserved and is visible only at the anterior end of the mold. The lower parts of the flanks are ornamented with fine riblets and striae which curve forward to the middle of the flanks, recurve sharply backward just above the middle, and then curve forward again. On the ventrolateral margins appear fairly prominent, widely spaced folds which curve strongly forward but do not cross the venter.

The figured specimen at a diameter of 36 mm. has a whorl height of 19.5 mm., a whorl thickness of 11 mm., and an umbilical width of 6 mm.

As far as preservation permits comparisons, this specimen seems to belong to *"Phylloceras" mazapilense* Burckhardt (1906, p. 125, Pl. 34, figs. 1-7, 19; 1930, Table 4, p. 50, 52, 69, 70) from the lower and upper Portlandian beds of the Mazapil region of Zacatecas, Mexico.

Figured Specimen: U. S. National Museum 103381.

Occurrence: Viñales limestone. Atlantic Refining Company of Cuba locality 5216.

Metahaploceras? sp.

(Plate 2, figures 8-10)

One silicified internal mold is placed provisionally in *Metahaploceras* rather than *Haploceras* or *Neochetoceras,* because its ribbing develops indistinct branching on the upper part of the flanks and is only gently falciform. It has a high and narrow whorl section, flattened flanks, narrowly rounded venter, and extremely narrow umbilicus. The indistinct prorsiradiate riblets and striae of the lower part of the flanks recurve gently backward just above the middle of the flanks and merge into numerous, stronger ribs that curve forward rather strongly on the upper part of the flanks and on the venter. The ribs attain their greatest strength on the ventrolateral shoulders, are nearly as wide as the interspaces, and tend to bifurcate on the upper third of the flanks.

The figured specimen, at a diameter of 32 mm., has a whorl height of 17.5 mm., a whorl thickness of 10 mm., and an umbilical width of 3 mm.

This species is much like *Metahaploceras mazapilense* (Burckhardt) (1906, p. 125, Pl. 34, figs. 1-7, 19) in whorl section but has a narrower umbilicus and does not develop widely spaced, prominent folds on the ventrolateral margins. Its ornamentation considerably resembles some of the coarser-ribbed species of *Haploceras,* such as *Haploceras costatum* Burckhardt (1906, p. 96, Pl. 25, figs. 1-10), but it is distinguished by the presence of rib-branching and a different whorl shape. On most species of *Haploceras* the venter is more broadly rounded, the ribbing is more falciform, and the place of greatest forward curvature of the ribs is lower on the flanks.

Figured Specimen: U. S. National Museum 103382.

Occurrence: Viñales limestone. Atlantic Refining Company of Cuba locality 7539.

Genus *Pseudolissoceras* Spath, 1925

Pseudolissoceras cf. *P. zitteli* (Burckhardt)

(Plate 4, figures 1-4, 7, 8, 11, 12)

This species is represented by more than 45 specimens, of which most are fragmentary, calcareous, internal molds. However, 18 specimens from Loma Sabanilla in Santa Clara Province are silicified. Form discoidal, compressed. Whorls subovate in section, higher than wide, thickest a little above the middle of the flanks, embracing about three fourths; flanks broad and nearly flat; venter evenly rounded on inner whorls, narrowly rounded on outer whorls. Umbilicus narrow, becoming wider during growth; wall low, vertical, rounding abruptly into flanks on outer whorls.

The whorls are marked with faint, gently falciform riblets and striae that are inflected forward most strongly near the middle of the flanks.

Dimensions in mm. are as follows.

Specimen	Greatest diameter	Whorl height	Whorl thickness	Umbilical width
Pl. 4, figs. 1–3	..	17	8 5	...
Pl. 4, figs. 7, 8	17	9	5	2 5
Pl. 4, figs. 11, 12	16	8	6	2 5

Some fragments show that the species attains a diameter of several inches.

The simple suture line is characteristic of *Pseudolissoceras* and contrasts markedly with the much more complicated suture line of *Haploceras*. The first lateral lobe is about twice as long as the siphonal lobe, is very broad, and bears a number of short, nearly equal-sized branches. The second lateral lobe is very short and small. Several auxiliary lobes are visible. Siphonal saddle broad, asymmetrically divided by a secondary lobe into two branches, of which the outer is higher. First lateral saddle about half the size of the siphonal saddles, asymmetrically divided.

The Cuban specimens are considered to represent a single species, although they vary somewhat in the relation of whorl height to thickness, in the degree of rounding of the upper part of the umbilical wall, and in the umbilical width. Similar variations noted in the described species of *Pseudolissoceras* have not been considered as worthy of varietal names. Most of the Cuban specimens correspond so closely with the figures of the Argentinian *P. zitteli* (Burckhardt) (1903, p. 55, Pl. 10, figs. 1-8; Haupt, 1907, p. 200, Pl. 7, figs. 3a, b, 4a-c; Krantz, 1928, p. 18, Pl. 1, fig. 6) that separation as a variety does not seem advisable. However, one of the specimens (Pl. 4, figs. 1-3) has a wider umbilicus than the others and possibly should be compared with *P. subrasile* (Burckhardt) (1906, p. 127, Pl. 34, figs. 8-14), from the lower Portlandian of Mexico, or to *P. rasile* (Oppel) (in Zittel, 1870, p. 173, Pl. 28, figs. 2, 3a-c), from the Portlandian of southern Europe. The Cuban specimens of *Pseudolissoceras* are associated with other ammonites of definite upper Portlandian age. In Mexico and Argentina the genus is known only from the lower Portlandian. European records indicate that the genus ranges through the Portlandian.

FIGURED SPECIMENS: U S National Museum 103383a, b, 103384a, b; 8 nonfigured specimens, U. S. National Museum 103385, 103386a-c, 103445a-d.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 268, 5216, 5229, 5231, 7539.

Genus *Hildoglochiceras* Spath, 1924

Hildoglochiceras cf. *H. grossicostatum* Imlay

(Plate 2, figures 3-5)

Two fragmentary internal molds retaining some shell material show the characteristic ornamentation of *Hildoglochiceras*. Whorls ovate in section, much higher than wide, thickest near middle of flanks, embracing most of preceding whorls; flanks broad and nearly flat, rounding evenly into venter, sloping gently toward umbilicus; venter narrowly rounded and keeled.

The whorls are ornamented with a prominent, narrowly rounded keel, sicklelike ribs and striae, and a weak spiral groove. The striae curve forward on the lower half of the flanks to the spiral groove, then backward to the upper third of the flanks, and then strongly forward. On the ventrolateral margins they pass into low, falcoid ribs which approach the keel at a sharp angle but generally flatten out before reaching it. The spiral groove is broad and shallow, occurs slightly above the middle of the flanks, and is bounded dorsally by a faint swelling.

This species is more like *Hildoglochiceras grossicostatum* Imlay (1939, p. 27, Pl. 2, figs. 5-11; Pl. 3, figs. 1-7, 9-11) than any other described species but has slightly finer ribbing. All the Indian species with similar ribbing are much less involute. Of these, *H. colei* Spath (1931, Pl. 99, fig. 6; Pl. 68, fig. 8, pl. 81, figs. 4a, b; Pl. 82, fig. 4) is the most similar.

Figured Specimens: U. S. National Museum 103387a, b.

Occurrence: Viñales limestone. Atlantic Refining Company of Cuba locality 5138.

Hildoglochiceras cf. *H. alamense* Imlay

(Plate 2, figures 1, 2, 6, 7)

Two specimens are probably identical specifically with *Hildoglochiceras alamense* Imlay (1939, p. 30, Pl. 4, figs. 6-9, 11, 12). The whorls are much higher than wide, the flanks are flattened, and the umbilicus is very narrow. The smaller specimen retains the shell and shows fine striae curved in the sigmoidal manner characteristic of the genus. Its venter is merely sharpened posteriorly but develops a low, rounded keel anteriorly. The larger specimen, preserved mainly as an internal mold, shows a well-developed spiral groove and traces of forwardly inclined ribbing on the venter. As the specimen is poorly preserved, the venter in most places appears to be merely sharpened, but there are indications that the shell bore a keel.

Dimensions in mm. are as follows:

Specimen	Greatest diameter	Whorl height	Whorl thickness	Umbilical width
Pl. 2, figs. 6, 7	22	11.5	5.5	2.5
Pl. 2, figs. 1, 2	43	23	12 (?)	6.5

Figured Specimens: U. S. National Museum 103388a, b.

Occurrence: Viñales limestone. Atlantic Refining Company of Cuba locality 5138.

Genus *Simoceras* Zittel, 1870

Simoceras sp. juv. cf. *S. volanense* (Oppel)

(Plate 3, figures 2, 3)

One small, silicified specimen is the sole representative of *Simoceras*. Whorls subquadrate in section, slightly wider than high, thickest near middle of flanks, barely embracing; flanks gently convex, becoming flatter anteriorly; venter flat and shouldered. Umbilicus very wide and shallow; wall very low, rounding evenly into flanks, becoming steeper anteriorly.

The ribs are broad, prominent, begin at the line of involution, incline forward gently on the flanks, and terminate in prominent tubercles on the ventrolateral shoulders. The ventral tubercles are connected by broad, weak ribs. The midventral line is nearly smooth.

At a diameter of 11 mm., the whorl height and thickness are each about 3 mm., and the umbilical width is 6 mm.

This specimen appears to be similar to the European *Simoceras volanense* (Oppel) (1863, p 231, Pl 58, figs 2a, b; Zittel, 1870, p 213, Pl. 32, figs 7-9), as well as the South American form *S.* aff *S. volanense* (Oppel) described by Krantz (1928. p 13, Pl. 3, fig 7).

FIGURED SPECIMEN: U. S. National Museum 103389.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba locality 268.

Genus *Virgatosimoceras* Spath, 1925

Virgatosimoceras? sp.

(Plate 3, figures 8-10)

One large fragment of an outer whorl retains considerable shell material Whorl subquadrate in section, as wide as high, thickest below middle, barely embracing; flanks gently convex; venter broadly rounded; umbilical wall low, steeply inclined, rounding rather abruptly into flanks

The shell is ornamented with strong ribs and minute radial striae. The ribs curve backward on the upper part of the umbilical wall, recurve rather abruptly low on the flanks, incline forward slightly on the flanks, and cross the venter transversely or with a gentle forward arching They are broad at their base, rounded or sharp-topped, of varying prominence, are weakest on the venter, and are widely but variably spaced. Some ribs remain simple, but most of them bifurcate on the upper third or fourth of the flanks. One rib bifurcates on the lower fourth of the flank, and both branches bifurcate again on the upper fourth. The thickness of the shell on the venter is about 5 mm.

The figured specimen at its anterior end has a whorl height and thickness of about 43 mm.

No described species of the Simoceratinae is very similar. *Virgatosimoceras albertinus* (Catullo) (*in* Zittel, 1870, p 222, Pl. 34, figs 1a-d) at a comparable size has a broader venter and stronger ribs of which all bifurcate low on the flanks. *Virgatosimoceras rothpletzi* (Schneid) (115, p. 88, Pl. 4, figs 1-1c; Pl. 7, figs. 2, 3-3b) has a similar whorl shape and type of rib branching, but its ribs are stronger and more widely spaced.

FIGURED SPECIMEN: U. S National Museum 103390.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba locality 5216.

Genus *Aspidoceras* Zittel, 1868

Aspidoceras spp.

(Plate 3, figures 4, 12, 13)

Two specimens from the Viñales limestone definitely belong to this genus. The larger specimen, preserved as a calcareous internal mold, has a depressed whorl section, a broad, gently rounded venter, and two closely associated rows of prominent spines The outer and larger row of spines marks the place of greatest whorl thickness The spines of the inner row are generally paired with those of the outer row and are on the upper margin of the umbilical wall. There is a suggestion of broad, ventral undulations connecting the spines.

The smaller specimen is only 9 5 mm in diameter, but it agrees in general form and ornamentation with the larger. As a result of silification the spines are excellently preserved and project nearly a mm. from the shell. At its anterior end the whorl height is 4 mm., and the whorl thickness is 6.5 mm., exclusive of the spines.

The larger specimen is comparable to the immature forms of the European *Aspidoceras rogosniciense* (Zeuschner) (*in* Zittel, 1868, p. 116, Pl. 24, figs. 4, 5-5d), although its spines are a little more closely spaced, and its ventral undulations are not nearly so strongly developed. The Mexican species, *A. alamitocensis* Castillo and Aguillera (1895, p. 43, Pl. 22, fig. 6; Pl. 23), is likewise very similar but probably has fewer spines per whorl

FIGURED SPECIMENS: U. S. National Museum 103391, 103392.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 268, 5216.

Genus *Physodoceras* Hyatt, 1900

Physodoceras sp.

(Plate 3, figure 11)

One fragment shows a form with a narrow umbilicus, a steep umbilical wall, broadly rounded flanks, several radially elongate umbilical tubercules, and extremely fine, closely spaced striae. The striae trend forward on the umbilical wall and on the lower part of the flanks but recurve near the middle of the flanks and are radial or rursiradiate on their upper part. The surface is slightly undulent but not ribbed.

The narrow umbilicus and single row of umbilical tubercles show that this species belongs in *Physodoceras* rather than *Aspidoceras*. *Physodoceras* is common in the Kimmeridgian and rare in the Portlandian. However, *P. cyclotus* (Oppel) (*in* Zittel, 1870, p. 201, Pl. 30, figs 2-5) is known from the Portlandian of Europe and a similar species (Krantz, 1928, p. 10, Pl. 1, fig. 1) from the Portlandian of South America.

FIGURED SPECIMEN: U. S. National Museum 103393.

OCCURRENCES: Viñales limestone. Atlantic Refining Company of Cuba locality 5216.

Genus *Virgatosphinctes* Uhlig, 1910

Virgatosphinctes cristóbalensis Imlay, n. sp.

(Plate 4, figure 13)

The species is represented definitely only by the type but probably includes several associated fragments of immature forms as well as a small specimen not from the type locality.

The outer whorl of the type has been smashed, but the penultimate whorl that appears to be undeformed is higher than wide, subquadrate in section, and thickest on the lower part of the flanks. The amount of involution is about two fifths. Flanks flattened, rounding rather abruptly into the moderately wide, flattened venter. Umbilicus moderate in width; wall low, vertical at base, rounding abruptly into flanks.

The ribbing of the inner whorls of the type specimen agrees, as far as observation permits, with that of the small specimen shown by Figures 5 and 6 of Plate 4. On the latter, the ribs are narrowly rounded and rather closely spaced. They begin at the line of involution, curve backward on the umbilical wall, incline forward strongly on the flanks, and cross the venter transversely without diminution in strength. Most of the ribs bifurcate on the upper third of the flank, but a few trifurcate, and a few remain simple. There are 5 or 6 narrow constrictions on each whorl The constrictions are generally preceded by a trifurcating or virgatomous rib and succeeded by a simple rib.

On the type specimen the ribbing of the penultimate whorl is like that just described, but the furcation points are a little lower. On the outer whorl the primary ribs become more widely spaced and are swollen on the umbilical shoulder into prominent bulges from which pass bundles of virgatomous ribs with as many as 5 to 8 ribs in each bundle. Branching occurs at various heights from the lower third to the upper third of the flanks. A constriction is followed by a simple rib, but all the other ribs are virgatomous. The suture line is not preserved.

The penultimate whorl of the holotype has a whorl height of 28 mm., and a whorl thickness of 20 mm. The small figured specimen (Pl. 4, figs. 5, 6), at a diameter of 44 mm., has a whorl height of 16 mm., a whorl thickness of 13 mm , and an umbilical width of 16 mm.

This species has remarkably similar ornamentation to that of *V. frequens* (Oppel) (1865, p. 295, Pl 87; Uhlig, 1910, p. 325, Pl. 63, figs 1a-c, 2, 3a-c; Pl. 75, figs. 1a-c; Pl. 75A, figs. 1a-c) but is less involute, has a more compressed whorl section, and develops thickened primary ribs at a much smaller size. A fragmentary specimen from the lower Portlandian of the Mazapil region of Mexico was described by Burckhardt (1906, p. 118, Pl. 32, fig. 2) as *Virgatites* sp. ind., but it is certainly a *Virgatosphinctes* similar to the Cuban species and to *V. frequens* (Oppel).

HOLOTYPE: U. S. National Museum 103394.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba locality 5229. *Virgatosphinctes* cf. *V. cristóbalensis* Imlay (U. S. National Museum 103395) was obtained at locality 5246.

Virgatosphinctes aff. *V. rotundidoma* Uhlig

(Plate 2, figure 14)

About 20 poorly preserved specimens represent an evolute, widely umbilicated species whose ribs are rather low, narrowly rounded, closely spaced, inclined slightly forward, and of which a little more than half bifurcate above the middle of the flanks. This species in form and ornamentation is similar to *V. rotundidoma* Uhlig (1910, p, 318, Pl. 52, figs. 1a-c; Pl. 53, fig. 1) but on the largest whorl has a greater frequency of single ribs and more ventrad furcation points. *Virgatosphinctes denseplicatus* (Waagen) (1875, p. 201, Pl. 46, figs. 3a-b; Pl. 55, figs. 1a, b; Spath, 1931. p. 532, Pl. 77, figs. 3a-c; Pl. 90, fig. 1, Pl. 96, figs. 3a, b; Pl. 102, fig. 4) has similarly dense ribbing in its inner whorls but develops thickened, distant primary ribs at an early stage and is less evolute.

FIGURED SPECIMEN: U. S. National Museum 103396.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 5216, 5229, 5231.

Genus *Corongoceras* Spath, 1925

Corongoceras filicostatum Imlay, n. sp.

(Plate 5, figures 1-16)

The species is represented by 35 silicified specimens of which most are fragments of inner whorls. Whorls subhexagonal in section, wider than high in early growth stages, becoming higher than wide in mature growth stages, thickest on lower part of flanks, embracing to the lateral tubercles, or about two fifths of the preceding volution; flanks evenly rounded in early stages, flattened in later stages; venter flattened and fairly narrow. Umbilicus fairly wide and shallow; wall low and rounding evenly into flanks.

The ornamentation consists of ribs and of lateral and ventral tubercles. The ribs begin at the line of involution, curve backward on the umbilical wall, incline forward moderately on the lower part of the flanks, recurve slightly on the upper part of the flanks, and cross the venter transversely. They are triangular in section, threadlike on the flanks but become broader above and are fairly coarse on the venter. On the innermost whorls to a diameter of about 15 mm., most of the ribs are simple. At greater diameters most of the ribs bifurcate or trifurcate a little above the middle of the flanks. Both simple and forked ribs bear lateral and ventral tubercles at diameters greater than about 5 mm. The lateral tubercles are small, acute, variable in size, and occur along a well-defined zone on the upper two fifths of the flanks. The ventral tubercles at all growth stages are larger and longer than the lateral tubercles and project laterally and outwardly. The ribs connecting the lateral with the ventral tubercles are commonly arranged in a zigzag fashion, but in places a pair of ribs issuing from a lateral tubercle terminate in a single ventral tubercle. Constrictions are pronounced on inner whorls.

The holotype (Pl. 5, figs. 12, 14, 15) at a diameter of 52 mm. has a whorl height of 19 mm., an estimated whorl thickness of 15 mm , and an umbilical width of 20 mm. The corresponding measurements of the paratype shown by Figures 5-7 of Plate 5 are 30 mm , 11 mm., 11 mm., and 12 mm. One specimen (Pl. 12, fig. 2) probably belonging to this species has a diameter three times greater than the holotype.

This species is placed in *Corongoceras* rather than *Protacanthodiscus* Spath (1923, p. 305): (1) because of the appearance of tubercles at an early age, (2) the uniform strength of the tubercles and their presence on all ribs, (3) the ribs branching high on the flanks, and (4) the absence of a smooth band on the venter.

Corongoceras filicostatum closely resembles *C. alternans* (Gerth) (1925, p. 89, Pl. 6, figs. 3, 3a) from the lower Tithonian of Argentina but may be distinguished by the greater regularity in strength of its ribs and tubercles. *Corongoceras mendozanum* (Behrendsen) (1891, p. 399, Pl. 25, figs. 2a-c), from slightly older beds, has coarser ribs and weaker tubercles. *Corongoceras lotenense* Spath (1925, p. 144; Haupt, 1907, p. 201, Pl. 9, figs. 7a-e) has wider-spaced ribbing and a broader venter, but the sculpture of its inner whorls is very similar.

TYPES: Holotype, U. S. National Museum 103397; 8 figured paratypes 103398, 103399, 103401a-f; 11 nonfigured paratypes 103402.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 248, 268, 5229, 7539.

Dickersonia Imlay, n. gen.

This genus has strongly tuberculate inner whorls that are nearly identical with those of *Corongoceras*, but its outermost whorl completely lacks tubercles and is ornamented only by bipartite ribbing similar to that of *Berriasella*. It is easily distinguished from *Berriasella* by its lack of umbilical tubercles at any stage of growth, by the presence of prominent lateral tubercles on the inner and intermediate whorls, and by the rather widely spaced rows of prominent ventral tubercles on the immature whorls. The simplification of the ornamentation during growth is exactly the opposite of that of *Blanfordiceras*, although the inner whorls of the two genera are somewhat similar in lateral aspect. *Protacanthodiscus* is distinguished from *Dickersonia* by the presence of umbilical tubercles, by the irregular occurrence of lateral tubercles, and by the increasing complexity of the ornamentation during growth.

Dickersonia is represented in the Cuban collections by two species, of which *D. sabanillensis* Imlay is designated as the genotype. Form discoidal, compressed, moderately evolute. Inner whorls subhexagonal in section, wider than high, strongly convex on flanks, flattened on venter. Outer whorls subovate in section, higher than wide, flattened on flanks, broadly rounded on venter. Umbilicus fairly wide. Umbilical wall fairly low, rounding abruptly into flanks on outer whorls Shell ornamented with simple and bifurcating ribs and with lateral and ventral tubercles. The ribs curve backward on umbilical wall, incline forward gently on flanks, and cross venter nearly transversely. On the inner whorls all ribs bear acute lateral tubercles and slightly stronger ventral tubercles, are highest and widest on the upper part of the flanks, and are reduced in strength on the venter. During growth the tubercles become less prominent, the ribs become higher and narrower, and the ventral thinning of the ribs becomes less pronounced. The largest outer whorls are not tuberculated and are ornamented by high, narrowly rounded ribs that are strongest on the venter. Accompanying the loss of tuberculation, the furcation points become less distinct, and intercalary ribs appear. Constrictions are deep and narrow, and are particularly pronounced on the inner whorls. The suture line is unknown.

No Mexican or South American species referable to *Dickersonia* have been described. Some of the immature ammonites from the Salt Range of India, placed by Spath (1939, p. 43-48, Pls. 4-6 in part) in *Blanfordiceras,* considerably resemble the immature forms, or inner whorls, of *Dickersonia* in lateral aspect but have an entirely different ventral aspect. However, one species described by Spath (1939, p. 47, Pl. 6, fig. 9; Pl. 18, figs. 7a, b; Pl. 20, figs. 7a, b) as *Blanfordiceras* (gen nov.?) sp. nov. possibly belongs to *Dickersonia* but is more coarsely ribbed than the Cuban species. It is associated with an ammonite assemblage which Spath (1939, p. 124, 125) considers Tithonian.

Dickersonia sabanillensis Imlay, n. sp.

(Plate 6, figures 9-12, 14-20)

This species is represented by 15 silicified, fragmentary specimens showing nearly all the growth stages. Whorls in early growth stages subhexagonal in section and wider than high, in later growth stages subovate in section and higher than wide, thickest near the middle of the flanks, embracing to the lateral tubercles or about two fifths of the preceding whorl. Flanks strongly convex in earliest growth stages; later becoming flattened and subparallel below the lateral tubercles but converging markedly above to the venter; in mature forms nearly flat below but rounding evenly above into venter. Venter flattened and moderately broad on innermost and on the intermediate-sized whorls, but broadly rounded on the outer whorls. Umbilicus fairly wide; wall low, rounding evenly into flanks on inner whorls, rounding abruptly on outer whorls.

The ornamentation consists of fairly closely spaced ribs and of lateral and ventral tubercles. The ribs begin at the line of involution, curve backward on the umbilical wall, incline forward moderately on the flanks, and cross the venter transversely. The character of the ribbing changes considerably with age On the inner whorls the ribs are threadlike on the lower part of the flanks, rather broad on the upper part of the flanks and on the venter, and are reduced in strength on the venter. During growth the ribs on both flanks and venter become high and narrow, and the ventral thinning becomes less pronounced. On the adult whorls the ribs are highest and strongest on the venter. On the innermost whorls about half of the ribs bifurcate, and on the outer whorls nearly all the ribs bifurcate along a well-defined zone a little above the middle of the flanks. All ribs bear small, acute tubercles of slightly variable size along the zone of furcation. All ribs terminate ventrally in acute tubercles that are nearly uniform in strength and are slightly larger than the lateral tubercles. Both lateral and ventral tubercles become less prominent during growth and are absent on the largest whorls. Concomitant with the disappearance of the tubercles the furcation points become less distinct, and some intercalary ribs appear. There are five or six deep, narrow constrictions on each whorl. Generally, the rib anterior to the constriction is somewhat larger than average Suture line unknown.

The holotype (Pl. 6, figs 14, 15) at a diameter of 41 mm. has a whorl height of 14 mm., a whorl thickness of about 15 mm, and an umbilical width of 17 mm The corresponding measurements of the paratype shown by Figures 11 and 12 of Plate 6 are 21 5 mm, 9 mm., 10 mm., and 8 5 mm.

This species differs from *Dickersonia ramonensis* Imlay by its weaker and more closely spaced ribs and tubercles on the inner whorls, thicker ribs on the outer whorls, the disappearance of tubercles at an earlier stage, and a slightly wider venter.

TYPES: Holotype, U. S. National Museum 103403; 5 figured paratypes, 103404a-e; 8 nonfigured paratypes 103405a-h.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 211(?), 215, 217, 268, 7539.

Dickersonia ramonensis Imlay, n sp

(Plate 6, figures 1-8)

This species is represented by 25 silicified specimens. Whorls in early stages subhexagonal in section and wider than high, in later stages subovate in section and higher than wide, thickest near middle of flanks, embracing about two fifths. Flanks in early stages rather strongly rounded below and converging markedly toward the venter, later flattened below and rounding evenly above. Venter flattened and fairly broad on innermost and intermediate-sized whorls, broadly rounded on outer whorls. Umbilicus fairly wide; wall fairly low, rounding evenly into flanks on inner whorls, rounding abruptly on outer whorls.

The ornamentation consists of moderately spaced ribs and of lateral and ventral tubercles. The ribs curve backward on the umbilical wall, incline gently forward on the flanks, and cross the venter transversely. On the inner whorls the ribs are fine on the lower part of the flanks; broad on the upper part of the flanks and on the venter; and are somewhat reduced in strength on the venter. During growth the ribs become high, narrow, more closely spaced, and the ventral thinning less pronounced. On the largest known whorl (Pl. 6, figs 2, 3) the ribs on the venter are excavated on their posterior sides and are not reduced in strength along the midline. On the inner whorls about half the ribs bifurcate, and on the outer whorls nearly all the ribs bifurcate slightly above the middle of the flanks. All ribs bear acute tubercles along the zone of furcation, and all ribs terminate ventrally in somewhat stronger tubercles. The tubercles are prominent on the inner whorls but decrease in strength during growth and are absent on the largest whorls. There are five or six deep, narrow constrictions on each whorl Suture line unknown

The holotype (Pl 6, fig. 8) at a diameter of 43 mm has a whorl height of about 18 mm. and an umbilical width of 15 mm. The paratype (Pl. 6, figs. 4, 5) at a diameter of 20 mm. has a whorl height of 7 mm., a whorl thickness of 9 mm, and an umbilical width of 9 mm.

The inner whorls of this species are intermediate in coarseness of ornamentation and width of section between those of *Dickersonia sabanillensis* Imlay and *Corongoceras filicostatum* Imlay. The outer whorls differ from those of *C. filicostatum* by complete lack of tuberculation and much more closely spaced ribbing. The specimen (U. S. National Museum 103410) shown in Figure 13 of Plate 6 is considerably corroded but appears to have finer and more widely spaced ribbing than is typical of *D. ramonensis*.

TYPES: Holotype, U. S. National Museum 103406; 4 figured paratypes, U. S. National Museum 103407a-d; 12 nonfigured paratypes, U. S. National Museum 103408a-i, 103409a-c.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 105(?), 268, 5231(?), 7539.

Genus *Micracanthoceras* Spath, 1925

Micracanthoceras sp. juv.

(Plate 4, figures 9, 10, 14-16)

The genus is represented by three small specimens which show several inner whorls. The specimen shown in Figures 14 and 16 of Plate 4 may be described as follows: whorls ovate in section, wider than high, embracing about one third; flanks evenly convex; venter broadly rounded. Umbilicus moderately wide; wall vertical at base, rounding evenly into flanks. Ribs high, narrow, strongly prorsiradiate on the flanks, nearly transverse on the venter. On the posterior part of the outer whorl most of the ribs are simple. On the anterior part about two thirds of the ribs bifurcate a little below the middle of the flanks. The bifurcation points are swollen but not tuberculate.

The specimen shown in Figures 9 and 10 of Plate 4 differs from that just described by its more depressed whorl section and by its slightly coarser, less forwardly inclined ribbing. The largest fragment (Pl. 4, fig. 15) has nearly straight ribs which bifurcate at the middle of the flanks and are slightly swollen at the points of furcation. The general appearance of these specimens is much like that of *Micracanthoceras alamense* Imlay (1939, p. 45, Pl. 9, figs. 3-12), from the Tithonian of Mexico.

FIGURED SPECIMEN: U. S. National Museum 103411a, b, 103412.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 268, 7539.

Genus *Durangites* Burckhardt, 1910

Durangites cf. *D. acanthicus* Burckhardt

(Plate 8, figures 5-7)

One small, silicified fragment of part of a whorl shows the essential features of the genus *Durangites*. Whorl ovate in section, slightly wider than high; flanks gently convex; venter nearly evenly rounded, slightly flattened along midventral line. Umbilical wall steeply inclined at base, rounding evenly into flanks.

The ornamentation consists of many high, narrow, straight, tuberculate ribs which are nearly radial on the umbilical wall and on the flanks, cross the venter transversely at the posterior end of the fragment but at the anterior end incline forward as a gentle sinus. Nearly all ribs bifurcate at or slightly below the middle of the flanks. The points of furcation are elevated but not tuberculated. All ribs terminate ventrally in radially compressed tubercles of varying prominence. Generally every other pair of secondary ribs terminates ventrally in a single tubercle of considerable prominence. Thus, four ventral tubercles correspond to three primary ribs, and every third ventral tubercle is particularly prominent. All ribs are much reduced in strength along the midventral line but not interrupted by a groove. One constriction present. The anterior end of the figured specimen has a whorl height of about 8.5 mm. and a whorl thickness of 9 mm.

This specimen closely resembles *Durangites acanthicus* Burckhardt (1912, p. 146, Pl. 36, figs. 7, 8, 10, 11, 15), particularly the posterior end of the outer whorl of Burckhardt's type but has fewer simple ribs and lacks lateral tubercles. As it is in the "Köllickeri stage" of development (Burckhardt, 1912, p. 144) it might be confused with species of *Micracanthoceras,* but that genus retains lateral tubercles to a much later growth stage, and at a comparable size is generally much more sparsely ribbed.

Figured Specimen: U. S. National Museum 103413.

Occurrence: Viñales limestone. Atlantic Refining Company of Cuba locality 268.

Durangites vulgaris Burckhardt

(Plate 3, figure 1)

1912. *Durangites vulgaris* Burckhardt, Inst. geol. México Bol. 29, p. 149, Pl. 37, figs. 1-36; Pl. 38, figs. 1-4.

One specimen showing only the lateral view of the outer whorl appears to be identical with the more convex variety of *Durangites vulgaris* Burckhardt from the upper Portlandian of México. Whorl strongly convex at its posterior end and gently convex at its anterior end. Umbilicus fairly wide; wall low and oblique, rounding evenly into flanks.

The ornamentation consists of high, narrow, tuberculate ribs that begin at the line of involution, curve forward to the middle of the flank, then recurve and incline slightly backward on the upper part of the flanks. On the posterior half of the outer whorl about one third of the ribs bifurcate, and on the anterior half the majority of the ribs bifurcate between the middle and the upper fourth of the flanks. The points of furcation are elevated but not tuberculated. On the posterior part of the whorl every second or third rib passes ventrally into prominent tubercles. On the anterior part of the whorl only about one rib in every six passes into a ventral tubercle. Generally the ventral tubercle marks the union of two secondary ribs issuing from a common primary rib. No constrictions present.

The figured specimen at a diameter of 32 mm. has a whorl height of 11 mm. and an umbilical width of 13 mm. The corresponding measurements of one of the Mexican specimens of *D. vulgaris* figured by Burckhardt are 31, 12, and 13 mm.

Besides *Durangites vulgaris,* the only other Mexican species similar to the Cuban form is *D. nodulatus* Burckhardt (1912, p. 155, Pl. 38, figs. 5-7, 9, 10), which has wider-spaced ribs and fewer single ribs.

Plesiotype: U. S. National Museum 103414.

Occurrence: Viñales limestone. U. S. Geological Survey Mesozoic locality 18581.

Genus *Lytohoplites* Spath, 1925

Lytohoplites caribbeanus Imlay, n. sp.

(Plate 7, figures 1-9)

This species is represented by 11 fragmentary, more or less distorted internal molds, of which some bear fragments of shell material. Whorls subovate in section, in early stages wider than high, in later stages apparently a little higher than wide, thickest near middle of flanks, embracing about one third; flanks convex on inner whorls, flattened on outer whorls; venter broadly rounded on inner whorls, regularly rounded on outer whorls Umbilicus fairly wide, shallow; wall low, inclined on inner whorls and rounding evenly into flanks, nearly vertical on outer whorls and rounding rather abruptly into flanks.

The ribs on the inner whorls are nearly straight, fairly low, narrowly rounded, separated by flat interspaces from two to three times as wide, incline forward gently on the umbilical wall and flanks, and do not bifurcate. During growth the ribs become flexuous, higher, broader at their base, narrower at their top, and more widely spaced. The ribs on the outer whorls are radial on the umbilical wall, are inflected forward on the lower part of the flanks, recurve slightly on the upper part of the flanks, and arch forward gently on the venter They are triangular in cross-section, very broad at their bases, acute at their summit, highest on the ventrolateral margins, and faintly reduced in strength along the midventral line. Almost all ribs are simple, only two examples of furcation at the base of the flanks being noted. The interspaces are broadly rounded. Constrictions are not evident, although some interspaces are deeper than others.

Most of the specimens are too distorted for accurate measurements, but the holotype appears to be only slightly compressed. At a diameter of 48 mm. it has a whorl height of 17 mm., a whorl thickness of 16(?) mm., and an umbilical width of 20 mm.

The specimens show some variation in density of ribbing but not enough to suggest specific differences. Gerth (1925, p. 99) has noted considerable variation in density of ribbing on *Lytohoplites burckhardti* (Mayer Eymar) in Burckhardt (1900, p. 17, Pl. 26, figs. 1, 2; 1903, p. 61, Pl. 10, figs. 17-20), but that species differs by the development of rather prominent ventral tubercles. Much more similar to the Cuban species is *L. vetustoides* (Burckhardt) (1903, p. 62, Pl. 10, figs. 23-25) which, however, may be distinguished by its slightly weaker ribbing and the development of a more definite ventral sinus. The form from Algiers described by Roman (1936, p. 31, Pl. 1, figs. 6, 6a) as *Himalayites ? abnormis* has similarly strong, simple ribs but has ventral tubercles and a well-defined ventral sinus. It probably should be referred to *Lytohoplites*.

TYPES: Holotype, U. S. National Museum 103415; 6 figured paratypes, U. S. National Museum 103416a-f.

OCCURRENCE: Viñales limestone Atlantic Refining Company of Cuba locality 5229.

Genus *Parodontoceras* Spath, 1923

Parodontoceras butti Imlay, n. sp.

(Plate 7, figures 10-12)

This species is represented by numerous specimens, but most of them are very poorly preserved. Whorl section subovate, higher than wide, embracing about one fourth; flanks flattened; venter narrowly rounded; umbilicus wide and shallow.

The ribs of the innermost whorls (Pl. 7, figs. 10, 12) are high, narrow, closely spaced, inclined forward on the flanks, transverse on the venter, and most of them bifurcate below the middle of the flanks During growth the points of furcation rise to, or a little above, the middle of the flanks, the anterior rib branches tend to separate as intercalary ribs, the frequency of simple, unbranched ribs increases, and the interspaces become wider. At a diameter greater than 50 mm. all the ribs are simple, strong, straight, widely separated, and only gently inclined forward. The ribs along the midventral line are not reduced in strength at any stage of development. Constrictions not present. Suture line unknown. Specimens too distorted for accurate measurements.

This species differs from *Parodontoceras antilleanum* Imlay, n. sp., with which it is associated, by being much more coarsely and sparsely ribbed at all stages of growth. That it is not an extreme variation of *P. antilleanum* is indicated by the lack of intermediate forms. The most similar foreign species is *Berriasella simplicicostata* Mazenot (1939, p. 135, Pl. 22, figs. 1, 2, 5a, b) from the Berriasian of France, which, however, has a slower rate of coiling and is more evolute. *Berriasella simplicicostata* is placed by Mazenot (1939, p. 127) in the group of *B. pontica* (Retowski), which Spath (1923, p. 305; 1925, p. 145) includes in his genus *Parodontoceras,* based on *Odontoceras callistoides* Behrendsen (1891, p. 402, Pl. 23, figs. 1a, b; Steuer, 1897, p. 41 [167], Pl. 17 [31], figs. 13-16; Burckhardt, 1906, p 139, Pl. 39, figs. 5, 6). The ribbing of the inner whorls of *P. butti* Imlay and *P. antilleanum* Imlay is nearly identical with that of *Parodontoceras callistoides* (Behrendsen) and *P. beneckei* (Steuer) (1897, p. 42, Pl. 17 [31], figs. 6-12) from the late Jurassic of South America, but the ribbing of the outer whorls differs by the complete lack of bifurcating ribs. The difference is probably only one of degree, as *Parodontoceras* characteristically has many simple and intercalary ribs on the outer whorls. It appears, therefore, that the Cuban species, along with *Berriasella simplicicostata* Mazenot, may be included under *Parodontoceras* as species showing extreme simplification of the ribs. The presence of a ventral sulcus on the South American species is confined mainly to the internal mold, as noted by Steuer (1897, p. 41), and, as in *Kossmatia,* is apparently slight or absent where the shell is preserved. It is not a basis for separating generically the American species of *Parodontoceras* from the European species of the group of *Berriasella pontica* (Retowski).

TYPES: Holotype, U. S. National Museum 103417; 2 figured paratypes, U. S. National Museum 103418a, b; 7 nonfigured paratypes, U. S. National Museum 103419a-g.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 5216, 5229, 5231.

Parodontoceras antilleanum Imlay, n. sp.

(Plate 8, figures 4-9)

This species is one of the most common in the Viñales limestone, but most specimens are much compressed and weathered. Whorl section subovate, higher than wide, except in earliest growth stages, embracing about one third; flanks flattened, venter narrowly rounded; umbilicus wide and shallow.

The ribs on the innermost whorls (Pl. 8, figs. 7-9) are fine, densely spaced, inclined forward on the flanks, transverse on the venter, and generally bifurcate below the middle of the flanks. During growth the points of bifurcation rise to near the middle of the flank, the anterior branches tend to separate as intercalary ribs, simple ribs become more common, and the interspaces become wider. At a diameter greater than about 50 mm. all the ribs are simple, moderately strong, nearly straight, fairly widely separated, and moderately inclined forward. The ribs along the midventral line are not reduced in strength at any stage of development. Constrictions not present. Suture line unknown. Most of the specimens are considerably compressed, but that shown in Figures 8 and 9 of Plate 8 has a whorl height of 17 mm. and a whorl thickness of 12 mm.

In fineness of ribbing this species resembles the small specimen figured by Burckhardt (1906, p. 139, Pl. 39, fig. 6) as *Hoplites* cf. *H. callistoides* Behrendsen from the Tithonian of Mexico, but at a comparable size has less flexuous ribs, of which fewer bifurcate.

TYPES. Holotype, U. S National Museum 103420; 4 figured paratypes, U. S. National Museum 103421a-d; 4 nonfigured paratypes, 103422a-d.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 249, 5216, 5229, 5231(?).

Genus *Berriasella* Uhlig, 1905

Berriasella? sp.

(Plate 8, figures 2, 3)

Two small fragments belong to a species with form and ribbing somewhat similar to *Berriasella lorioli* (Zittel) (1868, p. 103, Pl. 20, figs. 6, 8; Mazenot, 1939, p. 125, Pl. 19, figs. 3a-d, 4a, b, 5a, b, 6a, b, 7a, b). Whorl subquadrate in section, a little higher than wide, embracing about one third; flanks nearly flat; venter evenly rounded. Ribs high, narrow, radial, slightly flexuous, widely spaced, bifurcating regularly on the upper third of the flanks, and crossing the venter transversely without diminution in strength. Several intercalary ribs present on venter.

FIGURED SPECIMEN: U. S. National Museum 103423.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 268, 5216.

Genus *Spiticeras* Uhlig, 1903

Spiticeras? sp.

(Plate 8, figure 1)

One crushed, much weathered specimen was provisionally assigned by Dickerson to "*Astiera*" cf *A. astieriformis* Böse. The only distinct features are a series of prominent, rounded umbilical nodes from which radiate bundles of 4 or 5 narrowly rounded ribs that are closely spaced, incline slightly forward, and apparently do not branch on the flanks. The umbilicus appears to be fairly wide.

Generic determination of this specimen is not possible, but it probably belongs to the family Olcostephanidae, which would indicate an age not older than upper Portlandian

FIGURED SPECIMEN: U. S. National Museum 103424.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba locality 5271.

Genus *Leptoceras* Uhlig, 1883

Leptoceras? hondense Imlay, n. sp.

(Plate 10, figures 5-9, 11, 12)

This species is represented by dozens of specimens, of which most are much weathered and compressed. Coiling criocerold, whorls widely separated, ovate in section, higher than wide, increasing slowly in size. Flanks and venter evenly convex, flanks becoming flatter during growth.

First whorl beyond embryonic chambers apparently smooth. Succeeding whorls marked by simple, radial ribs that are strong on the flanks, considerably reduced in strength on the venter, and on the larger whorls bear small ventral tubercles. Interspaces from 1½ to 2 times wider than ribs, slightly wider relatively on the outer than on the inner whorls. No constrictions present.

Leptoceras? hondense Imlay is the most common uncoiled ammonite in the Cuban collection. Well-preserved specimens are in general easily distinguished from the associated *L. catalinense* Imlay by their coarser ribbing and slower rate of coiling. Many immature or weathered specimens cannot be placed definitely in either species, and some specimens appear to be intermediate between the two species. Among European species, *L. gracile* (Oppel) (Zittel, 1870, p. 233, Pl 36, figs. 3a-d), from the late Jurassic of southern Europe, is similar but has coarser and wider-spaced ribbing. Dickerson, in his preliminary studies of the fauna of the Viñales limestone, considered *L.? hondense* as probably identical with *L. beyrichii* (Karsten) (1858, p. 103, Pl. 1, figs. 4a-d; 1886, p. 61, Pl. 1, figs .4a-d), from the Barremian of Colombia, but the Cuban species may be distinguished by its sparser ribbing.

Leptoceras? hondense may be identical with *Ancyloceras* cf. *A. annulatus* D'Orbigny of Kellum (1937, p. 88, Pl. 9, figs. 1-3), from the upper Portlandian of the San Carlos Mountains of Tamaulipas, Mexico.

TYPES: Holotype, U. S. National Museum 103425; figured paratypes, U. S. National Museum 103426, 103427a-c; 6 nonfigured paratypes 103428a-f.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 257, 5216, 5246, 5272.

Leptoceras? catalinense Imlay, n. sp.

(Plate 10, figures 1-4)

The species is represented by about 25 specimens. Coiling crioceroid. Whorls moderately to fairly widely separated, ovate in section, higher than wide, increasing rapidly in size. Flanks and venter evenly convex.

First whorl beyond embryonic chambers apparently smooth. Succeeding whorls marked by simple, radial, or slightly prorsiradiate ribs that thicken ventrally, and on the larger whorls bear small ventral tubercles. Interspaces equal to, or slightly wider than the ribs.

This species shows considerable variation in the rate of coiling, but the outer whorls are generally not as widely separated as in *L. hondense* Imlay, and the ribbing is much denser. *Crioceras* sp. ind. Burckhardt (1919, p. 58; 1921, Pl. 21, fig. 3), from the *Substeueroceras* beds of Mexico, is probably a similar species, but its much larger size precludes close comparison. *Leptoceras* sp. ind. Mazenot (1939, p. 245, Pl. 40, fig. 2), from the Berriasian of France, has sparser ribbing that is slightly rursiradiate, but its general appearance is very similar. The resemblances of *L.? catalinense* and *L.? hondense* to some of the Barremian species (Uhlig, 1883, p. 146-150, Pls. 29 and 32 in part) of Europe is rather remarkable and undoubtedly influenced former assignments of the Viñales limestone to an early Cretaceous age

TYPES: Holotype, U. S. National Museum 103429; figured paratypes, U. S. National Museum 103430a, b, 103431, 103432.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 257, 5216, 5271(?), 5272.

Genus *Hamulina* D'Orbigny, 1850

Hamulina? rosariensis Imlay, n. sp.

(Plate 9, figures 1-11; plate 12, figure 1)

This species is represented by numerous straight and hook-shaped fragments, but none shows the complete form, and nearly all are much weathered. It cannot be ascertained whether the species has 2 or 3 limbs, but the latter seems more probable. The small, coiled embryonic portion is shown in Figure 2 of Plate 9 The hook-shaped fragments have one long, nearly straight limb that passes by means of a 180° turn into a much shorter, subparallel limb. The whorl section is ovate and perhaps slightly higher than wide.

The ribs are strong, rounded, nontuberculated, incline forward gently on the flanks but cross the venter and dorsum transversely, are a little stronger on the venter than on the dorsum, and increase regularly in strength during growth. The interspaces on the posterior part of the longer limb are about as wide as the ribs and anteriorly become somewhat wider. On the crook the interspaces are about twice as wide as the ribs, and on the shorter limb are about three times as wide as the ribs. There is no reduction of ribbing along the midline of the venter or dorsum.

This species cannot be placed satisfactorily in any of the genera, *Bochianites*, *Leptoceras*, or *Protancyloceras*, to which have been referred the few uncoiled ammonites found until now in late Jurassic deposits. Although agreeing in general form and ornamentation with *Hamites* of the Albian stage it may be distinguished by its ribbing, stronger on the dorsum and wider-spaced anteriorly. Besides, the great difference in age makes generic identity with *Hamites* seem very doubtful. Certain of the nontuberculate species of *Hamulina* (Gignoux, 1920, p. 129) of the Barremian stage are likewise similar, although not as coarsely ribbed. Assignment of the Cuban species to a new genus is considered inadvisable, because of the poor preservation of the fossils and lack of knowledge of the suture line.

TYPES: Holotype, U. S National Museum 103433; 8 figured slabs of 28 paratypes, U. S. National Museum 103434a-f, 103435a, b.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 5216, 5229, 5231.

Genus *Ptychoceras* D'Orbigny, 1842

Ptychoceras? sp.

(Plate 10, figure 10)

One internal mold shows parts of two appressed limbs ornamented with straight, strong, narrowly rounded ribs that become strongest ventrally. The ribs incline gently forward on the smaller limb and gently backward on the larger limb. The interspaces on the smaller limb are a little narrower than the ribs, but on the larger limb become a little wider. The mold resembles *Ptychoceras* in form and ornamentation but is possibly distinguished by its large size and by the strength of the ribbing on the smaller limb. As *Ptychoceras* has been recorded only from the Barremian, Aptian, and Albian stages of the Lower Cretaceous, it undoubtedly did not range down into the Upper Jurassic. The Cuban specimen is too poorly preserved to serve as the type of a new genus but is worthy of notice as another example of the rich and varied assemblage of uncoiled ammonites that characterized the late Jurassic deposits of Cuba in marked contrast to the apparent paucity of uncoiled ammonites of the same age in other parts of the world.

FIGURED SPECIMEN: U S National Museum 103436

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba locality 184.

"Genus" *Lamellaptychus* Trauth, 1927

Lamellaptychus rectecostatus (Peters) emend. Trauth?

Four immature valves from Cuba were questionably referred by Trauth (1936, p. 69) to *Lamellaptychus rectecostatus* (Oppel, 1863, Pl. 70, figs. 1, 4), which he said was characterized by the straightness of the imbricating ribs on the convex side of the valve, although in the earliest stages the ribs may be slightly flexuous. The specimens examined by Trauth are from Soroa in Pinar del Río Province and are associated on the same rock with *L. angulocostatus* (Peters) and *L. seranonis* (Coquand). None of the specimens in the present collection seems referable to this species. According to Trauth, *L. rectecostatus* is common in the late Upper Jurassic of the Alpine-Mediterranean region but rare is the early Lower Cretaceous.

Lamellaptychus angulocostatus (Peters)

(Plate 11, figures 8-10)

For synonymy see TRAUTH: K. Akad. Wetensch. Amsterdam, Sec. Sci. Pr., vol. 39, no. 1, p. 70, 1936.

Trauth considers that most of the specimens described by O'Connell (1921, p. 10-12, Figs. 15, 16, 18, not Fig. 17) as *Aptychus pimientensis* belong to *L angulocostatus,* which is common in the Neocomian of the Alpine-Mediterranean regions and is characterized by the acute angle formed by the ribs on the convex side of the valve. The collections in hand contain 23 specimens which agree in form and sculpture with those figured by O'Connell from Cuba and also with those figured by Pictet and Loriol (1858, p. 46-48, Pl. 10, figs. 3, 5a-c, 6a-c, 7, 8a-d, 9, 10-12) from the Neocomian of Switzerland.

PLESIOTYPES: U. S. National Museum 103437, U. S. National Museum 103438, U S National Museum 103439.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 142, 215, 237, 250, 257, 5271, 3083. O'Connell records this species from Mount Pimiento, 5 miles north of San Cristóbal and from Río Hondo, 7 miles northeast of San Cristóbal in Pinar del Río Province. Trauth records the species from Bahia Hondo and Soroa in Pinar del Río Province, and from between Encrucijada and Santa Clara in Santa Clara Province.

Lamellaptychus angulocostatus (Peters) var. *longa* Trauth

For synonymy see TRAUTH: K Akad. Wetensch. Amsterdam, Sec. Sci. Pr., vol. 39, no. 1, p. 71, 1936.

Trauth distinguishes this variety from the typical form by its much more slender form and notes that in the Alpine-Mediterranean region it is always confined to the Neocomian. None of the specimens available to the writer are nearly so slender as the type (Pictet and Loriol, 1858, p. 46-48, Pl. 10, fig. 4) of the variety, but Trauth records three specimens from Cuba, including one figured by O'Connell (1921, p. 11, 12, Fig. 17).

OCCURRENCES: Viñales limestone. O'Connell records the species from Río Hondo, 7 miles northeast of San Cristóbal in Pinar del Río Province. Trauth records it from between Encrucijada and Santa Clara in Santa Clara Province.

Lamellaptychus angulocostatus (Peters) var. *atlanticus* (Hennig)

1913. *Aptychus atlanticus* HENNIG, Deutsche geol. Gesell. Zeitschr, Band 65, p. 155, Pl. 2, figs. 1a-b.

1936. *Lamellaptychus angulocostatus* (Peters) var. *atlantica* (Hennig). TRAUTH, K. Akad. Wetensch. Amsterdam, Sec. Sci. Pr, vol 39, no. 1, p. 72.

Trauth referred four Cuban specimens to this variety, in which the convex side of the valve bears ribs that apically are acute-angled as in the typical form of *Lamellaptychus angulocostatus,* but that toward the external margin change to broad, backwardly directed curves. These curves are slightly offset apically at one or more places. The type of the variety was obtained in the Cape Verde Islands from limestones which Hennig considered Upper Jurassic, but Trauth suggested that they are probably Neocomian in age No specimens referable to this variety are present in the collections under examination.

Occurrences: Viñales limestone. Trauth records this species from between Encrucijada and Santa Clara in Santa Clara Province, and from Soroa and the region south of Toro in Pinar del Río Province.

Lamellaptychus angulocostatus (Peters) var. *cristóbalensis* (O'Connell)

(Plate 11, figures 2, 3)

1921. *Aptychus cristóbalensis* O'Connell, Am. Mus. Novit., no. 28, p. 7, 8, Figs. 7, 8.
1936. *Lamellaptychus angulocostatus* (Peters) var. *cristóbalensis* (O'Connell). Trauth, K. Akad. Wetensch. Amsterdam, Sec. Sci. Pr., vol. 39, no. 1, p. 73.

This variety, represented by seven specimens, is characterized by: (1) a rather pronounced keel extending from the apex diagonally to the junction of the lateral and external margins; (2) the acute-angled ribs near the apex; (3) the broad, backwardly curved ribs near the external margin; and (4) the more or less distinct offsets apically of the ribs in the vicinity of the keel. It is similar to the variety *atlantica* (Hennig) but may be distinguished by the extension of the acute-angled ribs farther toward the external margin and by the more convex lateral margins.

Plesiotypes: U. S. National Museum 103440a, b.

Occurrences: Viñales limestone. Atlantic Refining Company of Cuba localities 142 and 3083. O'Connell listed this species from the Río Hondo, 7 miles northeast of San Cristóbal and from the Río San Cristóbal on the Finca of Rafael Begoa, about 9 miles north of San Cristóbal in Pinar del Río Province. The variety has also been found by the writer about 4 miles southeast of La Peña, in southern Coahuila, Mexico, where it is associated with ammonites of lower Hauterivian age.

Lamellaptychus seramonis (Coquand)

(Plate 11, figures 4, 6)

For synonymy see Trauth: K. Akad. Wetensch. Amsterdam, Sec. Sci. Pr., vol. 39, no. 1, p. 73.

This is the most common aptychus in the available Cuban collections, being represented by 20 specimens O'Connell (1921, p. 9, 10, Figs. 9-14) studied 8 specimens, which she placed in a new species, *Aptychus cubanensis* O'Connell, but Trauth, who examined 2 Cuban specimens of the same species, proclaimed its identity with *L. seranonis* (Coquand). Judging from the illustrations by Pictet and Loriol (1858, p. 48-50, Pl 11, figs, 1, 2; 4, 7, 8), the determination by Trauth seems to be correct.

Lamellaptychus seranonis is characterized by: (1) an inward bending of the lateral margin in conformity with the course of the ribs; (2) a prominent keel extending from the apex diagonally to the junction of the lateral and external margins; (3) the ribs being broadly curved rather than angulated toward the external margin; and (4) the ribs bending gently apically as they approach the harmonic margin and concomitantly becoming larger, broader, flatter, and wider-spaced. It is distinguished from *L. angulocostatus* (Peters) by the inflection of the lateral margin and by the absence of angulated ribbing toward the external margin.

According to Trauth (1936, p. 74) *Lamellaptychus seranonis* (Coquand) is fairly common in the lower Neocomian and rare in the late Jurassic of the Alpine-Mediterranean region.

PLESIOTYPES: U. S. National Museum 103441; U. S. National Museum 103442.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba localities 127, 142, 251, 252, 5211, and U. S. Geological Survey Mesozoic locality 18579. O'Connell records this species as *Aptychus cubanensis* from 9 miles north of San Cristóbal in Pinar del Río Province. Trauth records it from Soroa in the same province.

Lamellaptychus excavatus Trauth

(Plate 11, figures 1, 5)

1936. *Lamellaptychus excavatus* TRAUTH, K. Akad. Wetensch. Amsterdam, Sec. Sci. Pr., vol. 39, no. 1, p. 73.

This species, represented by 13 specimens, has not been reported previously from Cuba, but appears to be identical specifically with the types (Pictet and Loriol, 1858, p. 48-50, Pl. 11, figs. 3, 5, 6) from the early Cretaceous of Switzerland. It is similar to *L. seranonis* (Coquand) in form and sculpture, but its ribs are more closely spaced throughout and do not bend apically as they approach the harmonic margin.

PLESIOTYPES: U. S. National Museum 103443a, b.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba locality 252.

Genus *Hadrocheilus* Till, 1907

Hadrocheilus sp.

(Plate 11, figure 7)

One cephalopod beak from the Viñales limestone apparently belongs to *Hadrocheilus* Till (1907, p. 560, 568, 659, 667, 675) and probably to the *Depressi* group. Only the dorsal surface is exposed, but this shows a short, much depressed form with a median and two lateral wings. The cape is depressed, broadly rounded apically, and bears a low, rounded dorsal keel. The shaft has elevated, sharp, lateral margins that are bounded inwardly by depressed areas, is marked medially by a small furrow, and is not distinctly impressed with respect to the cape.

The Cuban specimen appears to have a more depressed form and more elevated shaft margins than any described species of *Hadrocheilus* but is not sufficiently well preserved to merit a specific name. The genus, according to Till (1907, p. 675-680), ranges through the Jurassic and Lower Cretaceous but attains its greatest development in species and individuals in the early Lower Cretaceous. He indicates that the *Depressi* group of *Hadrocheilus* is confined to the Upper Jurassic and Neocomian.

FIGURED SPECIMEN: U. S. National Museum 103444.

OCCURRENCE: Viñales limestone. Atlantic Refining Company of Cuba locality 226.

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PLATE 1.—LATE JURASSIC FOSSILS FROM CUBA

Figure

Phylloceras pinarense Imlay, n. sp.

(1, 4) Lateral and ventral views of paratype U. S. Nat. Mus. 103379a (× 2).

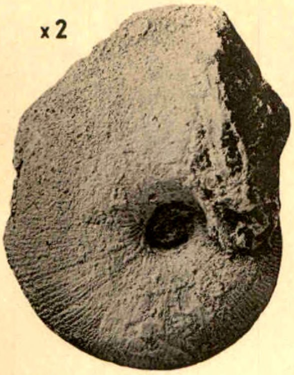
(2, 3) Lateral and apertural views of paratype U S. Nat. Mus. 103379b (× 2).

(5) Paratype U. S. Nat. Mus. 103379c.

(6-9) Suture, lateral view, cross section, and ventral view of holotype U. S. Nat. Mus. 103378.

All specimens from Atlantic Refining Company of Cuba locality 5216 (p. 1442).

Figures natural size unless otherwise indicated.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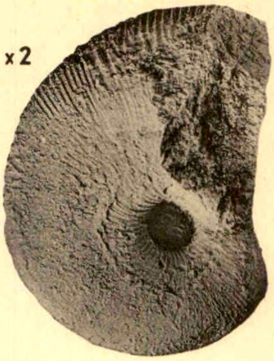
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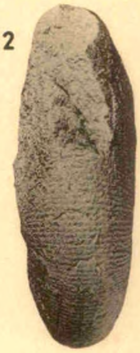
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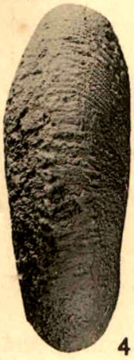
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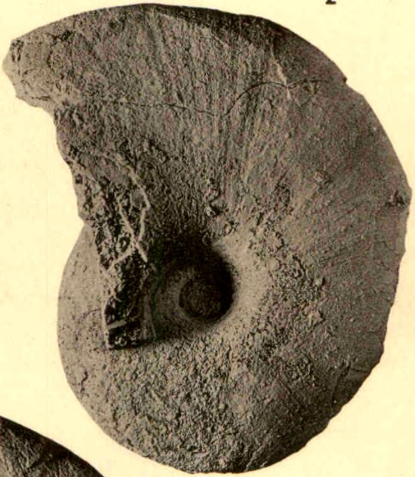
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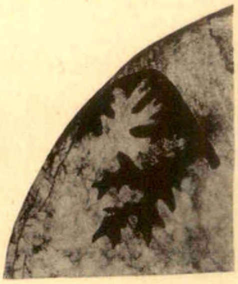
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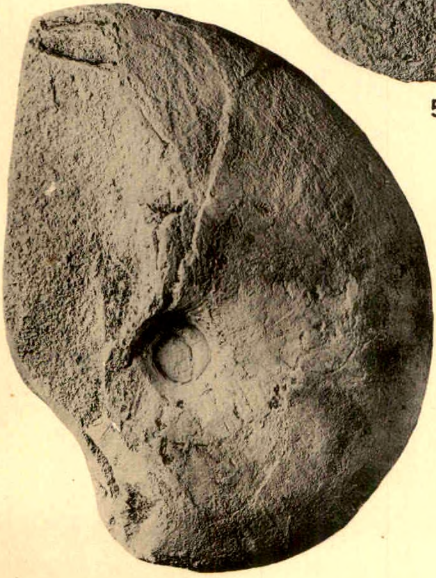
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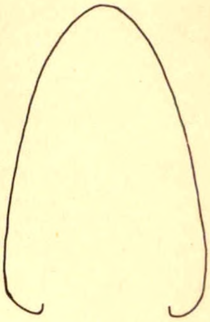
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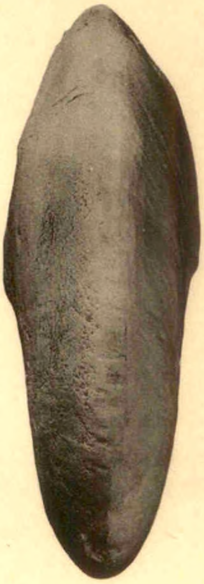
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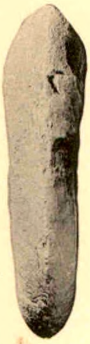
LATE JURASSIC FOSSILS FROM CUBA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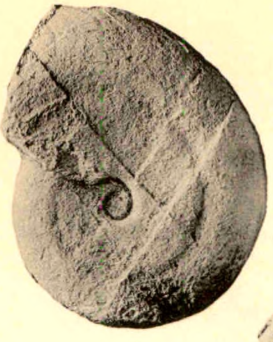
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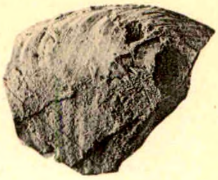
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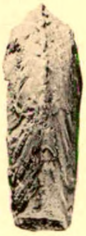
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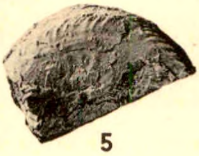
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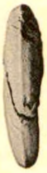
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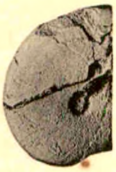
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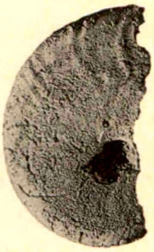
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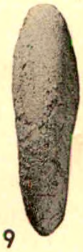
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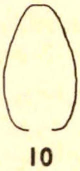
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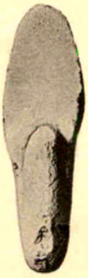
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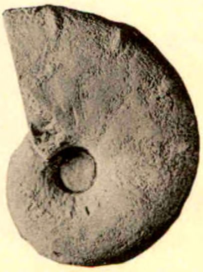
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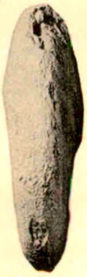
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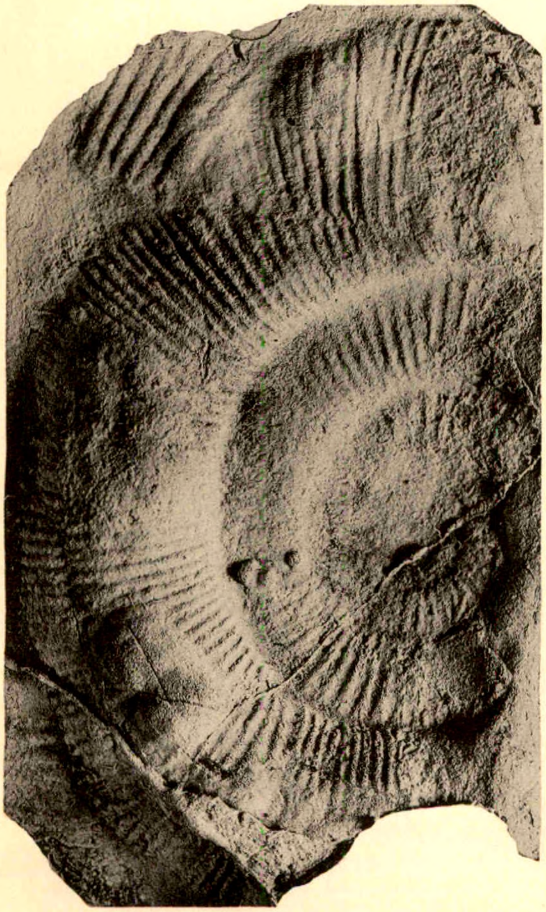
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LATE JURASSIC FOSSILS FROM CUBA

PLATE 2.—LATE JURASSIC FOSSILS FROM CUBA

Hildoglochiceras cf. *H. alamense* Imlay.

Figure

(1, 2) Ventral and lateral views of specimen U. S. Nat. Mus. 103388a.

(6, 7) Ventral and lateral views of specimen U. S. Nat. Mus. 103388b.
Both specimens from Atlantic Refining Company of Cuba locality 5138 (p. 1445).

Hildoglochiceras cf. *H. grossicostatum* Imlay.

(3, 4) Lateral and ventral views of specimen U. S. Nat. Mus. 103387a.

(5) Lateral view of specimen U. S. Nat. Mus. 103387b.
Both specimens from Atlantic Refining Company of Cuba locality 5138 (p. 1444).

(8-10) *Metahaploceras?* sp.

Lateral and ventral views and cross section of specimen U. S. Nat. Mus. 103382. Atlantic Refining Company of Cuba locality 7539 (p. 1443).

(11-13) *Metahaploceras* cf. *M. mazapilense* (Burckhardt).

Apertural, lateral, and ventral views of specimen U. S. Nat. Mus. 103381. Atlantic Refining Company of Cuba locality 6216 (p. 1442).

(14) *Virgatosphinctes* aff. *V. rotundidoma* Uhlig.

Specimen U. S. Nat. Mus. 103396. Atlantic Refining Company of Cuba locality 5216 (p. 1448).

All figures natural size.

PLATE 3.—LATE JURASSIC FOSSILS FROM CUBA

Figure

(1) *Durangites vulgaris* Burckhardt.

Plesiotype U. S. Nat. Mus. 103414. U. S. Geol. Survey Mesozoic locality 18581 (p. 1453).

(2, 3) *Simoceras* sp. juv. cf. *S. volanense* (Oppel).

Specimen U. S. Nat. Mus. 103389 (× 3). Atlantic Refining Company of Cuba locality 268 (p. 1445).

(4) *Aspidoceras* sp.

Specimen U. S. Nat. Mus. 103391. Atlantic Refining Company of Cuba locality 5216 (p. 1446).

(5-7) *Durangites* cf. *D. acanthicus* Burckhardt.

Lateral and ventral views of specimen U. S. Nat. Mus. 103413. Atlantic Refining Company of Cuba locality 268 (p. 1452).

(8-10) *Virgatosimoceras?* sp.

Lateral view, cross section, and ventral view of specimen U. S. Nat. Mus. 103390. Atlantic Refining Company of Cuba locality 5216 (p. 1446).

(11) *Physodoceras* sp.

Lateral view of specimen U. S. Nat. Mus. 103393. Atlantic Refining Company of Cuba locality 5216 (p. 1447).

(12, 13) *Aspidoceras* sp.

Ventral and lateral views of specimen U. S. Nat. Mus. 103392 (×3). Atlantic Refining Company of Cuba locality 268 (p. 1446).

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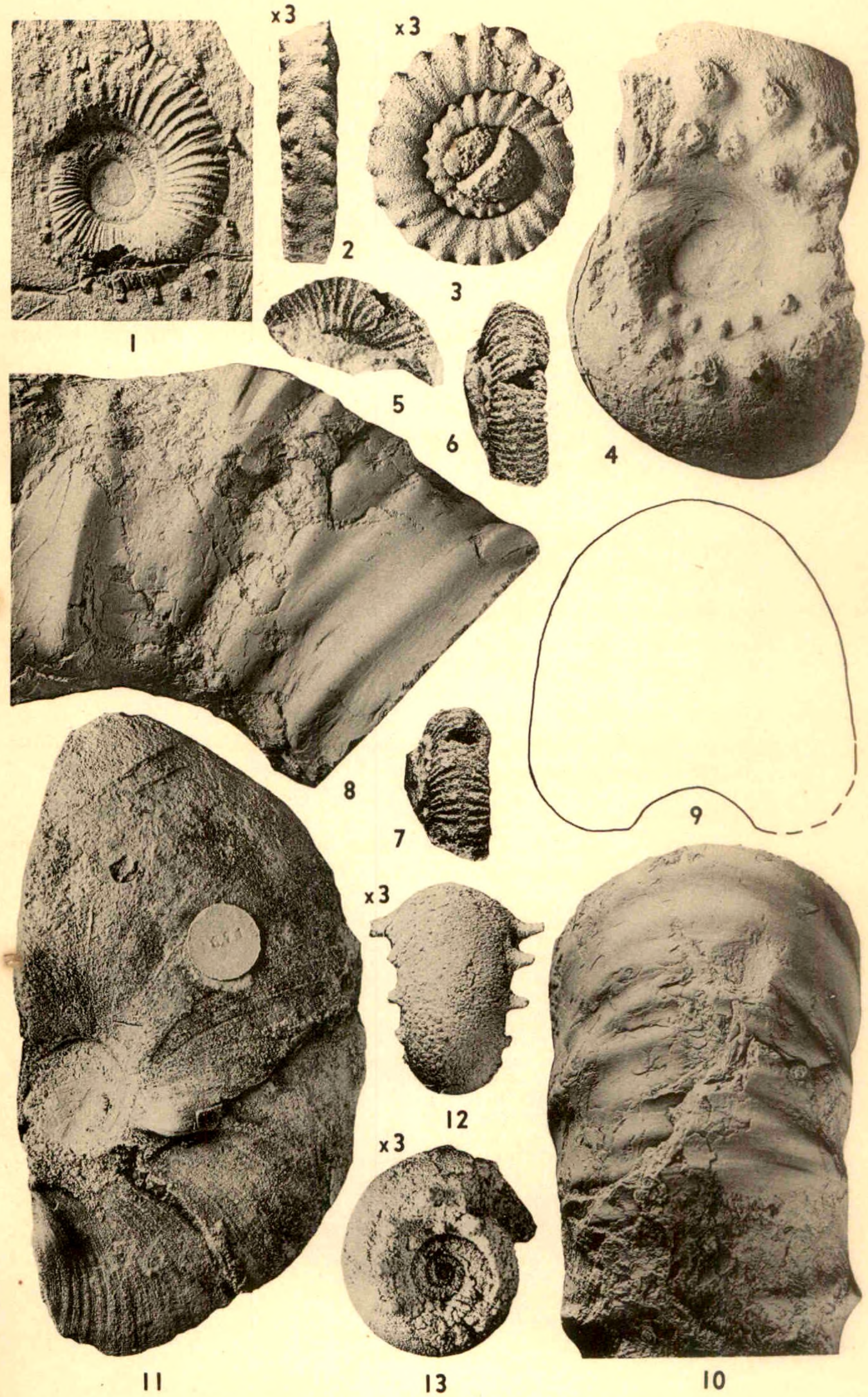
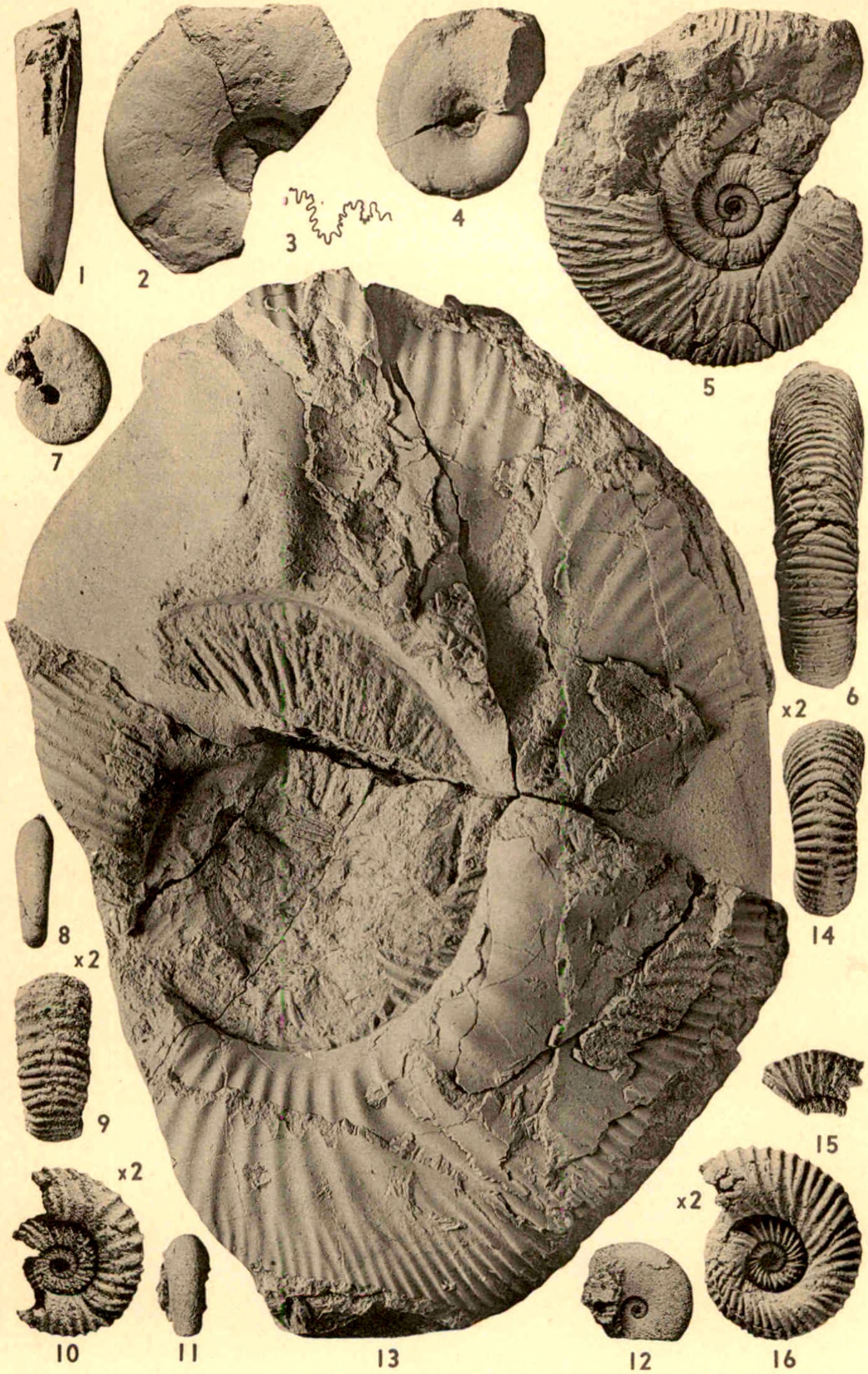


LATE JURASSIC FOSSILS FROM CUBA



LATE JURASSIC FOSSILS FROM CUBA

PLATE 4.—LATE JURASSIC FOSSILS FROM CUBA.

Pseudolissoceras cf. *P. zitteli* (Burckhardt).

Figure

(1-3) Views and suture of specimen U. S. Nat. Mus. 103383a.

(4) Lateral view of specimen U. S. Nat. Mus. 103383b.

(7, 8) Views of specimen U. S. Nat. Mus. 103384a.

(11, 12) Views of specimen U. S. Nat. Mus. 103384b.

Specimens represented by figures 1-4 are from the Atlantic Refining Company of Cuba locality 5231. Specimens represented by figures 7, 8, 11, and 12 are from Atlantic Refining Company of Cuba locality 268 (p. 1443).

(5, 6) *Virgatosphinctes* cf. *V. cristóbalensis* Imlay.

Views of specimen U. S. Nat. Mus. 103395 from Atlantic Refining Company of Cuba locality 5246 (p. 1448).

(13) *Virgatosphinctes cristóbalensis* Imlay, n. sp.

Holotype U. S. Nat. Mus. 103394 from Atlantic Refining Company of Cuba locality 5229 (p. 1447).

Micracanthoceras sp. juv.

(9, 10) Views of specimen U. S. Nat. Mus. 103411a (× 2).

(14, 16) Views of specimen U. S. Nat. Mus. 103411b (× 2).

(15) Lateral view of specimen U. S. Nat. Mus. 103412.

Specimens represented by figures 9, 10, 14, and 16 are from Atlantic Refining Company of Cuba locality 7539. Specimen represented by figure 15 is from the same company's locality 268 (p. 1452).

Figures natural size unless otherwise indicated.

PLATE 5.—LATE JURASSIC FOSSILS FROM CUBA

Corongoceras filicostatum Imlay, n. sp.

Figure

(1, 2) Lateral and apertural views of paratype U. S. Nat. Mus. 103401a.

(3) Paratype U. S. Nat. Mus. 103401b.

(4) Paratype U. S. Nat. Mus. 103401c.

(5-7) Paratype U. S. Nat. Mus. 103398.

(8) Suture of paratype, U. S. Nat. Mus. 103401d (× 2).

(9) Paratype U. S. Nat. Mus. 103401e.

(10, 11) Ventral and lateral views of paratype U. S. Nat. Mus. 103399.

(12, 14, 15) Holotype U. S. Nat. Mus. 103397, ventral view shows posterior part of outer whorl.

(13, 16) Paratype U. S. Nat. Mus. 103401f.

Specimen represented by figures 10 and 11 is from locality 248. Specimens represented by figures 5-7, 12, 14, and 15 are from locality 268. Remainder of specimens are from locality 7539 of the Atlantic Refining Company of Cuba (p. 1448).

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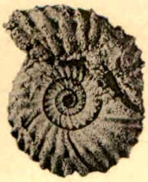

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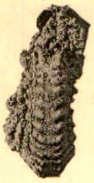

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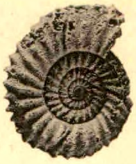

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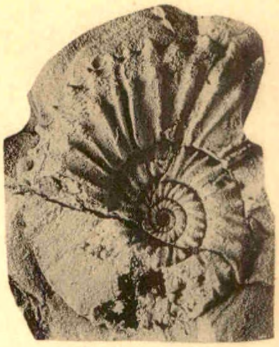

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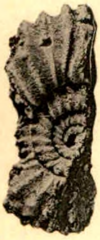

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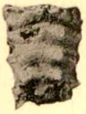

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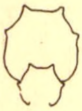

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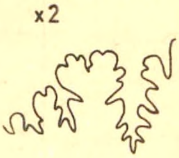

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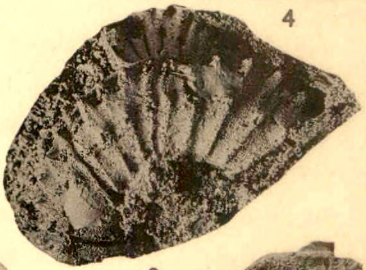

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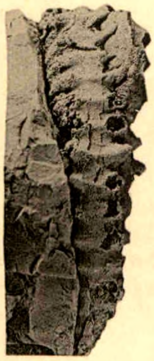

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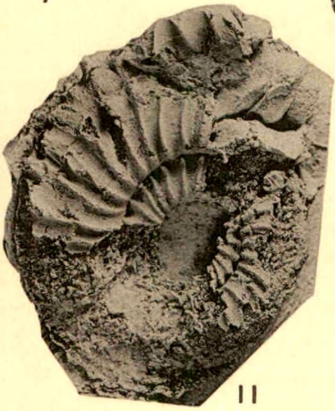

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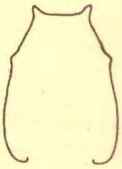

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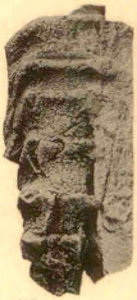

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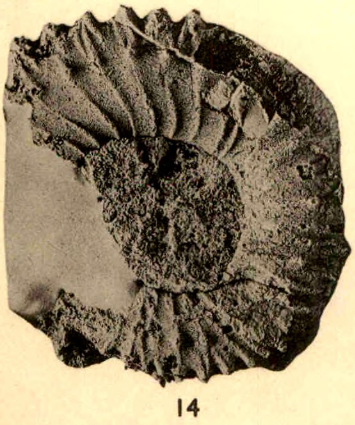

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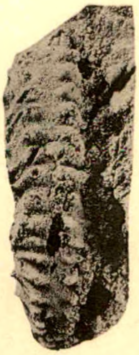

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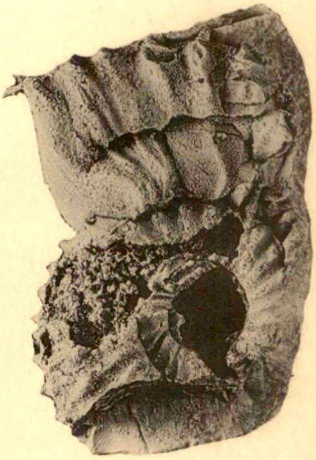

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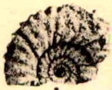
LATE JURASSIC FOSSILS FROM CUBA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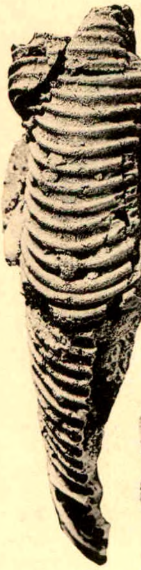
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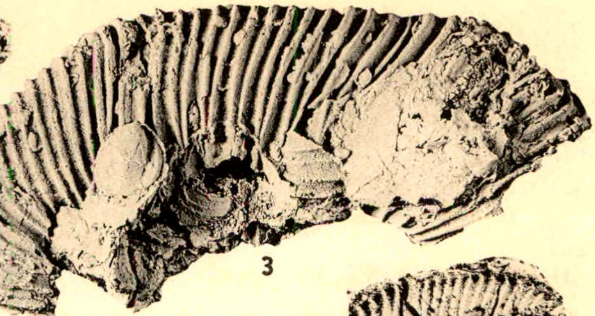
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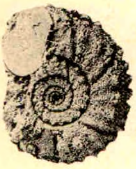
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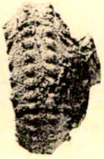
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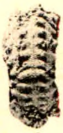
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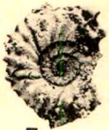
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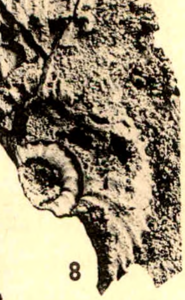
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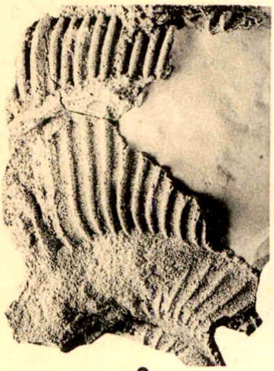
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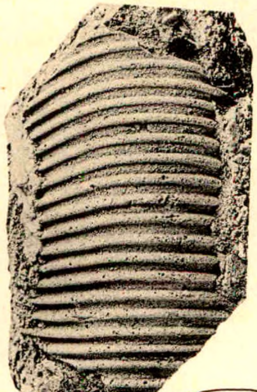
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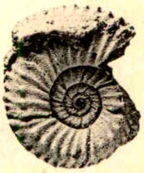
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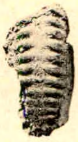
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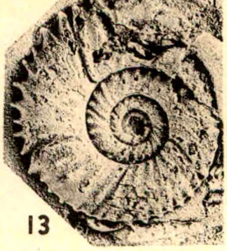
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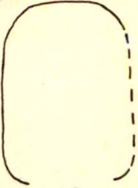
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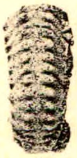
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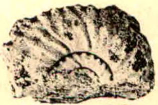
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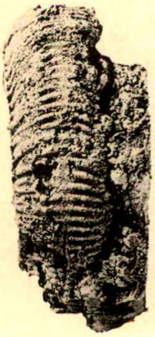
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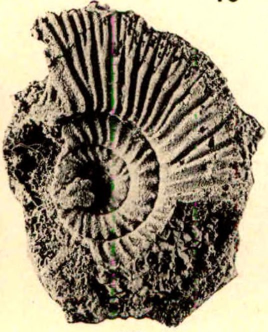
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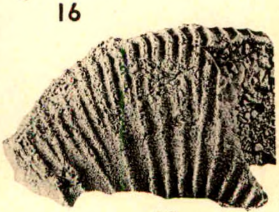
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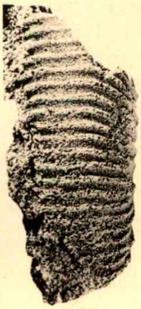
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LATE JURASSIC FOSSILS FROM CUBA

PLATE 6.—LATE JURASSIC FOSSILS FROM CUBA

Dickersonia ramonensis Imlay, n. sp.

Figure

(1) Paratype U. S. Nat. Mus. 103407a.

(2, 3) Ventral and lateral views of paratype U. S. Nat. Mus. 103407b, anterior part of ventral view directed downward.

(4, 5) Paratype U. S. Nat. Mus. 103407c.

(6, 7) Paratype U. S. Nat. Mus. 103407d.

(8) Holotype U. S. Nat. Mus. 103406.

All specimens from Atlantic Refining Company of Cuba locality 268 (p. 1451).

Dickersonia sabanillensis Imlay, n. sp.

(9) Lateral view of paratype U. S. Nat. Mus. 103404a.

(10) Ventral view of paratype U. S. Nat. Mus. 103404b.

(11, 12) Paratype U. S. Nat. Mus. 103404c.

(14, 15) Ventral and lateral views of holotype U. S. Nat. Mus. 103403.

(16, 19, 20) Cross section, lateral and ventral views of paratype U. S. Nat. Mus. 103404d.

(17, 18) Paratype U. S. Nat. Mus. 103404e.

All specimens from Atlantic Refining Company of Cuba locality 268 (p. 1450).

(13) *Dickersonia* cf. *D. ramonensis* Imlay

Specimen U. S. Nat. Mus. 103410 from Atlantic Refining Company of Cuba locality 7539 (p. 1452).

All figures natural size.

PLATE 7.—LATE JURASSIC FOSSILS FROM CUBA

Lytohoplites caribbeanus Imlay, n. sp.

Figure

(1, 2) Ventral and lateral views of holotype U. S. Nat. Mus. 103415.

(3) Paratype U. S. Nat. Mus. 103416a.

(4) Deformed paratype U. S. Nat. Mus. 103416b showing lateral and part of ventral views.

(5, 6) Cross section and lateroventral view of paratype U. S. Nat. Mus. 103416c showing ventral furrow.

(7) Paratype U. S. Nat. Mus. 103416d.

(8) Paratype U. S. Nat. Mus. 103416e.

(9) Paratype U. S. Nat. Mus. 103416f.

All specimens from Atlantic Refining Company of Cuba locality 5229 (p. 1453).

Parodontoceras butti Imlay, n. sp.

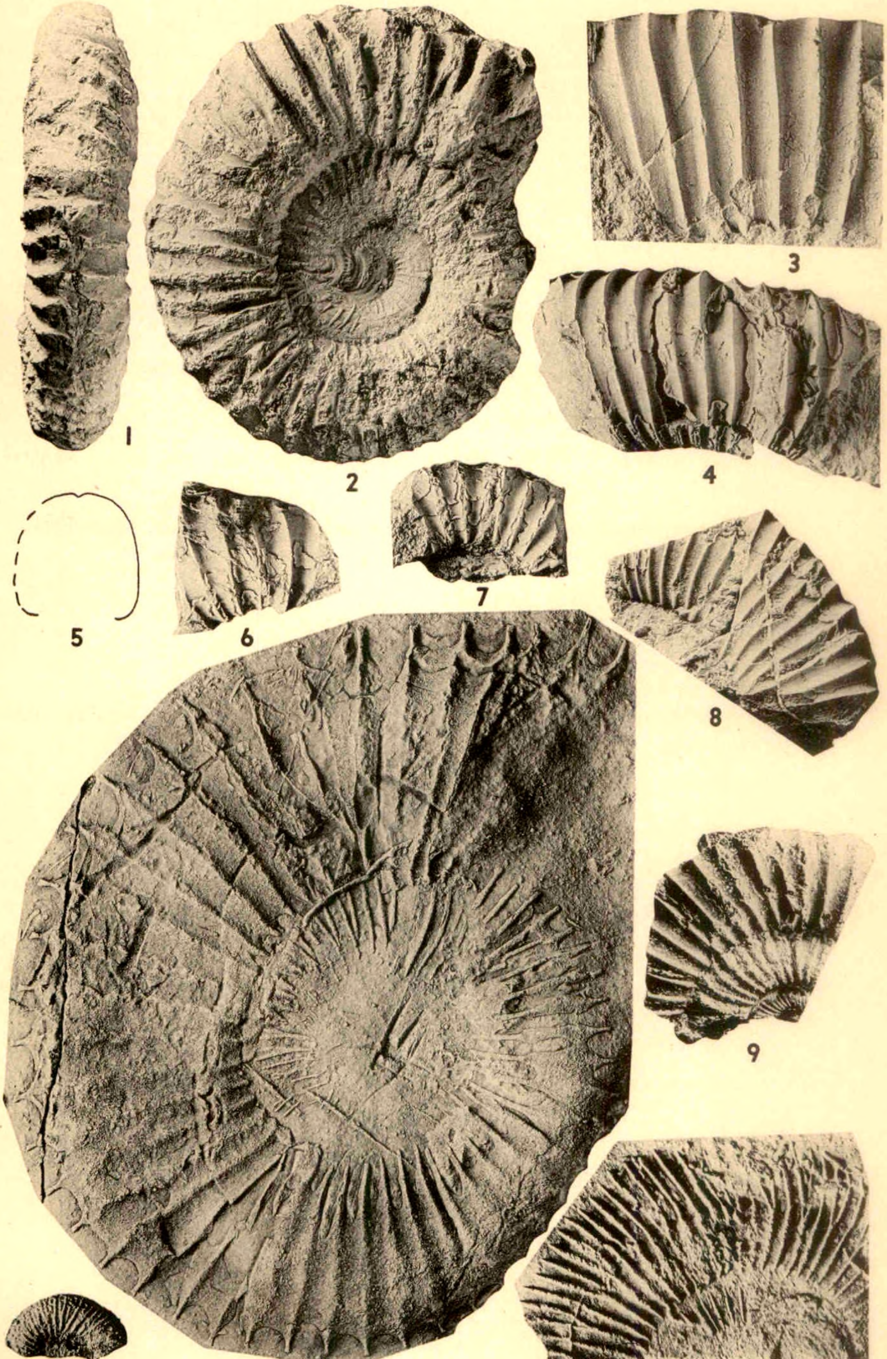
(10) Paratype U. S. Nat. Mus. 103418a.

(11) Holotype U. S. Nat. Mus. 103417.

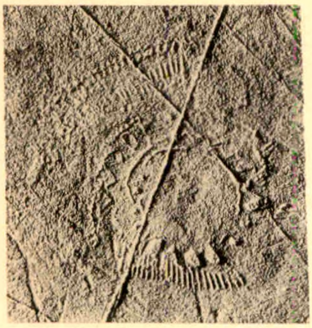
(12) Paratype U. S. Nat. Mus. 103418b.

All specimens from Atlantic Refining Company of Cuba locality 5216 (p. 1454).

All figures natural size.



LATE JURASSIC FOSSILS FROM CUBA



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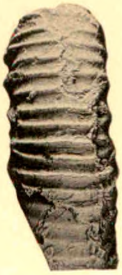

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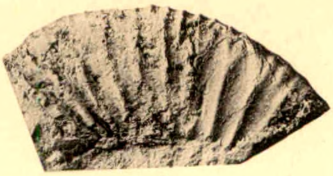

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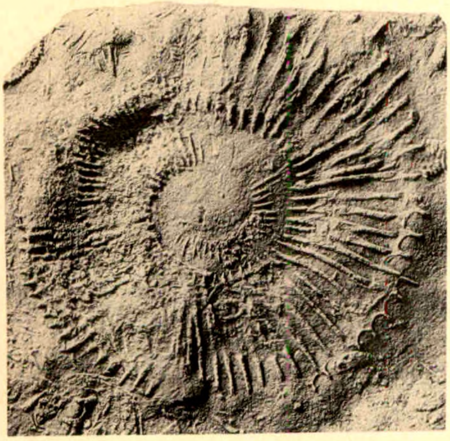

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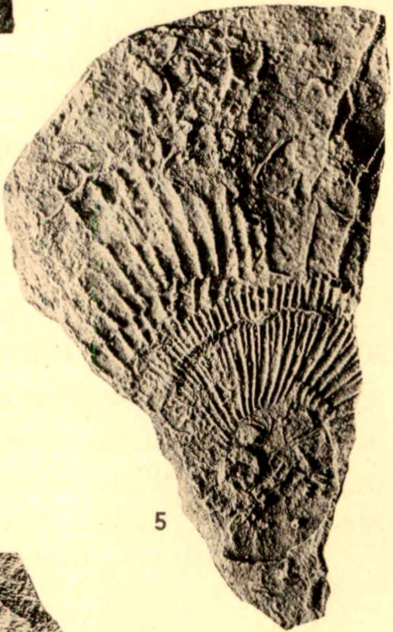

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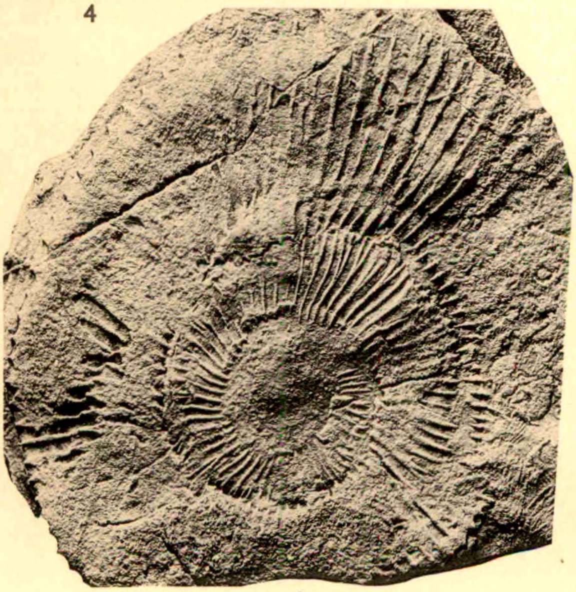

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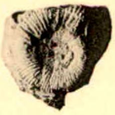

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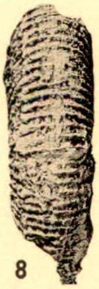

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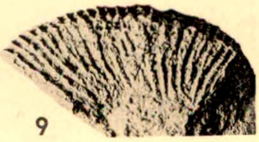

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LATE JURASSIC FOSSILS FROM CUBA

PLATE 8.—LATE JURASSIC FOSSILS FROM CUBA

Figure

(1) *Spiticeras?* sp.

Specimen U. S. Nat. Mus. 103424 from Atlantic Refining Company of Cuba locality 5271 (p. 1456).

(2, 3) *Berriasella?* sp.

Specimen U. S. Nat. Mus. 103423 from Atlantic Refining Company of Cuba locality 5216 (p. 1456).

Parodontoceras antilleanum, Imlay, n. sp.

(4) Paratype U. S. Nat. Mus. 103421a.

(5) Paratype U. S. Nat. Mus. 103421b.

(6) Holotype U. S. Nat. Mus. 103420.

(7) Paratype U. S. Nat. Mus. 103421c.

(8, 9) Paratype U. S. Nat. Mus. 103421d.

All specimens from Atlantic Refining Company of Cuba locality 5216 (p. 1455).

All figures natural size.

PLATE 9.—LATE JURASSIC FOSSILS FROM CUBA

Figure

Hamulina? rosariensis Imlay, n. sp.

(1) Four paratypes on slab U. S. Nat. Mus. 103435a.

(2) Three paratypes on slab U. S. Nat. Mus. 103434a, specimen on left has initial whorls.

(3, 5) Two paratypes on slab U. S. Nat. Mus. 103434b.

(4) Paratype U. S. Nat. Mus. 103434c.

(6-8) Three paratypes on slab U. S. Nat. Mus. 103435b.

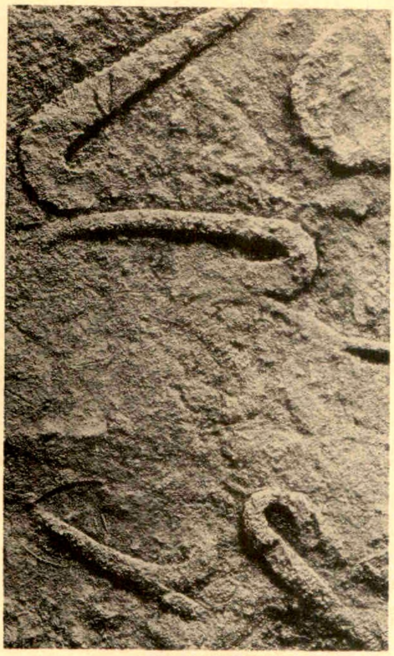
(9) Holotype U. S. Nat. Mus. 103433.

(10) Eight paratypes on slab 103434d.

(11) Four paratypes on slab 103434e.

Slabs 103435a, b are from Atlantic Refining Company of Cuba locality 5216.
Slabs 103434a–e are from Atlantic Refining Company of Cuba locality 5229 (p. 1457).

All figures twice natural size.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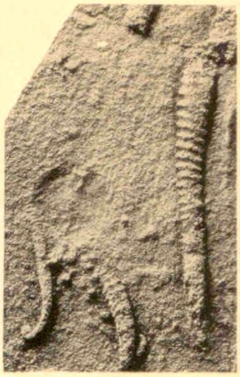
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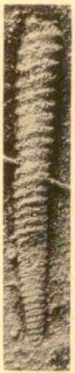
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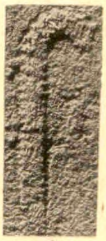
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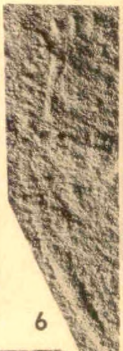
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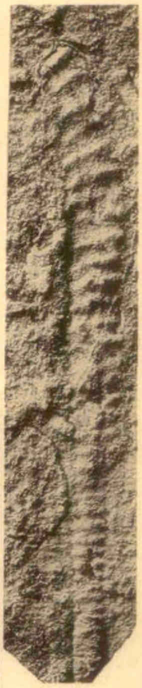
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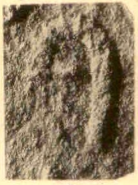
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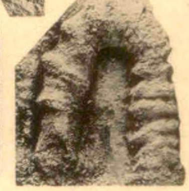
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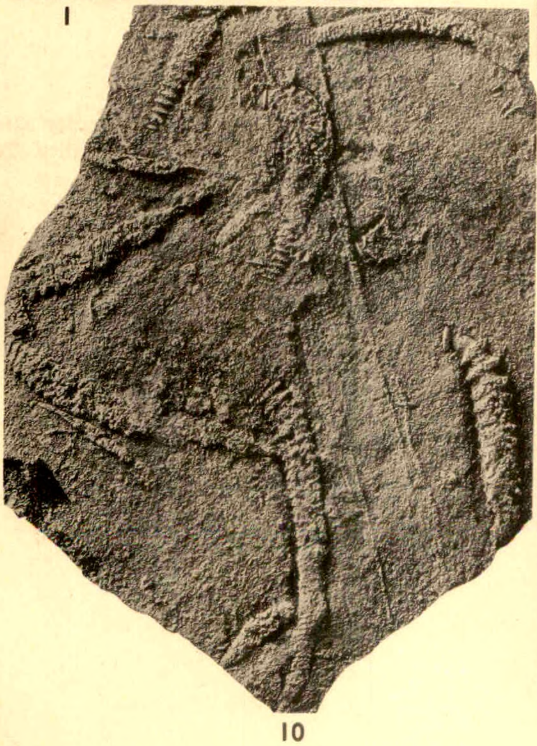
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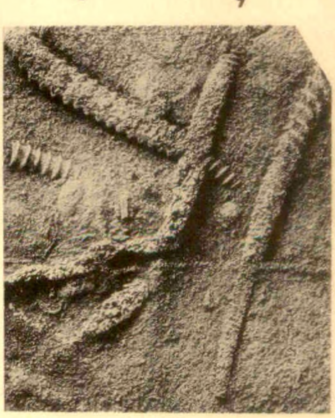
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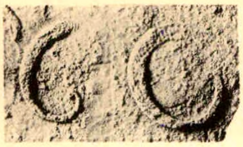
LATE JURASSIC FOSSILS FROM CUBA



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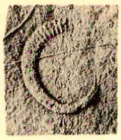

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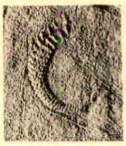

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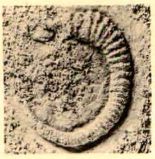

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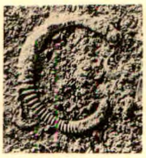

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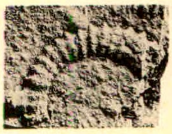

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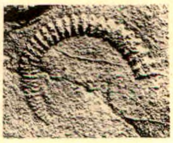

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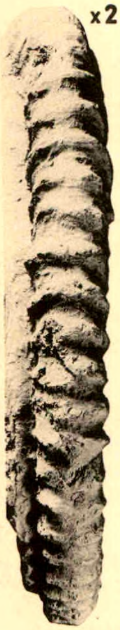

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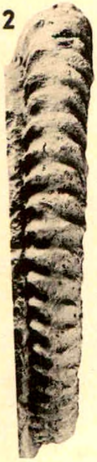

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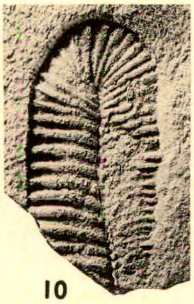

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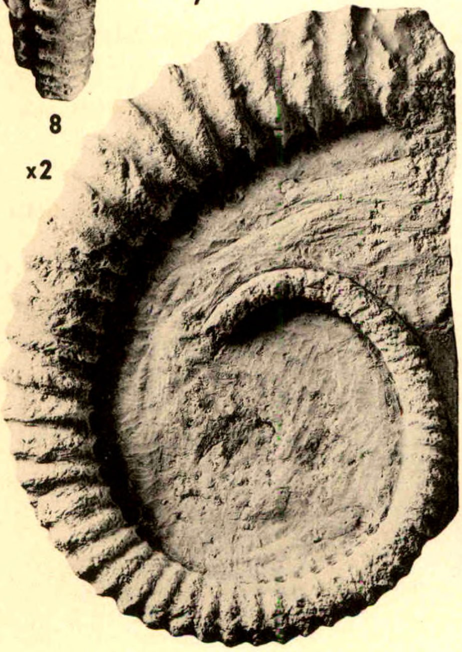
x2



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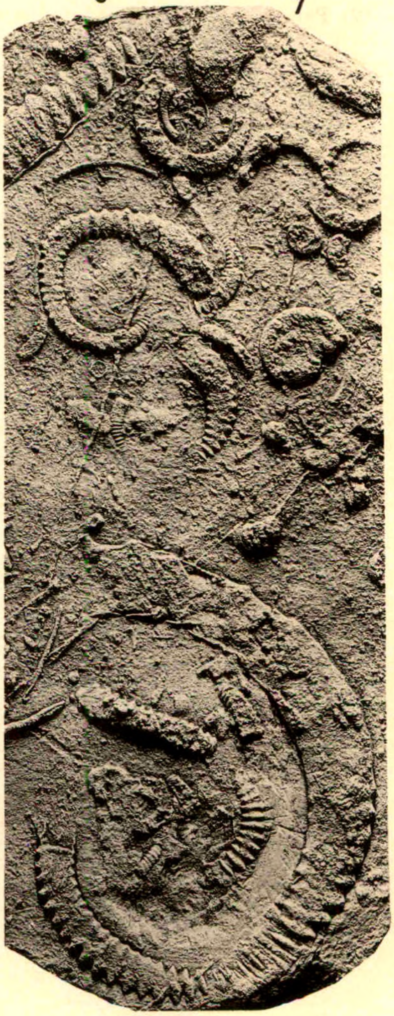


12

LATE JURASSIC FOSSILS FROM CUBA

Plate 10.—Late Jurassic Fossils from Cuba

Figure

Leptoceras? catalinense Imlay, n. sp.

(1) Two paratypes on slab U. S. Nat. Mus. 103430, specimen on left showing initial whorls, Atlantic Refining Company of Cuba locality 5216.

(2) Paratype U. S. Nat. Mus. 103432 from Atlantic Refining Company of Cuba locality 257.

(3) Paratype U. S. Nat. Mus. 103431 from Atlantic Refining Company of Cuba locality 5272.

(4) Holotype U. S. Nat. Mus. 103429 from Atlantic Refining Company of Cuba 5216 (p. 1457).

Leptoceras? hondense Imlay n. sp.

(5, 6) Paratypes U. S. Nat. Mus. 103427a, b from Atlantic Refining Company of Cuba locality 5216.

(7) Paratype U. S. Nat. Mus. 103426 from Atlantic Refining Company of Cuba locality 257.

(8, 9, 11) Ventral and lateral views of holotype U. S. Nat. Mus. 103425 (× 2) from Atlantic Refining Company of Cuba locality 5246.

(12) Paratypes on slab U. S. Nat. Mus. 103427c from Atlantic Refining Company of Cuba locality 5216 (p. 1456).

Ptychoceras? sp.

(10) Specimen U. S. Nat. Mus. 103436 from Atlantic Refining Company of Cuba locality 184 (p. 1458).

Figures natural size unless otherwise indicated.

Plate 11.—Late Jurassic Fossils from Cuba

Figure

Lamellaptychus excavatus Trauth

(1, 5) Plesiotypes U. S. Nat. Mus. 103443a, b from Atlantic Refining Company of Cuba locality 252 (p. 1461).

Lamellaptychus angulocostatus var. *cristóbalensis* (O'Connell)

(2, 3) Plesiotypes U. S. Nat. Mus. 103440a, b from Atlantic Refining Company of Cuba locality 142 (p. 1460).

Lamellaptychus seranonis (Coquand)

(4) Plesiotype U. S. Nat. Mus. 103441 from Atlantic Refining Company of Cuba locality 252.

(6) Plesiotype U. S. Nat. Mus. 103442 from U. S. Geol. Survey Mesozoic locality 18579 (p. 1460).

Hadrocheilus sp.

(7) Specimen U. S. Nat. Mus. 103444 from Atlantic Refining Company of Cuba locality 226 (p. 1461).

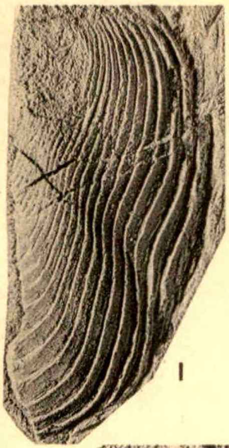
Lamellaptychus angulocostatus (Peters)

(8) Plesiotype U. S. Nat. Mus. 103438 from Atlantic Refining Company of Cuba locality 250.

(9) Plesiotype U. S. Nat. Mus. 103439 from Atlantic Refining Company of Cuba locality 3083.

(10) Pesiotype U. S. Nat. Mus. 103437 from Atlantic Refining Company of Cuba locality 142 (p. 1459).

All figures twice natural size.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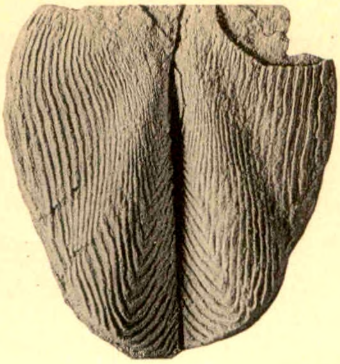
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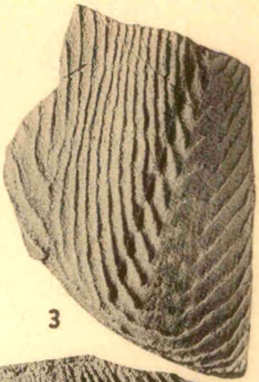
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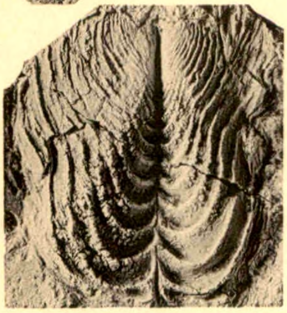
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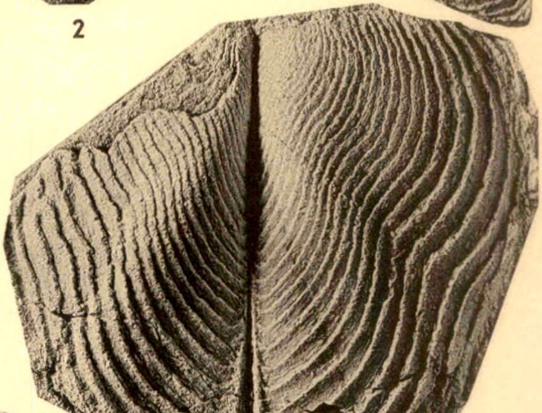
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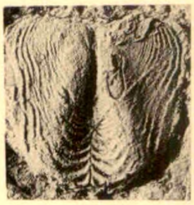
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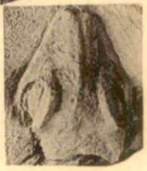
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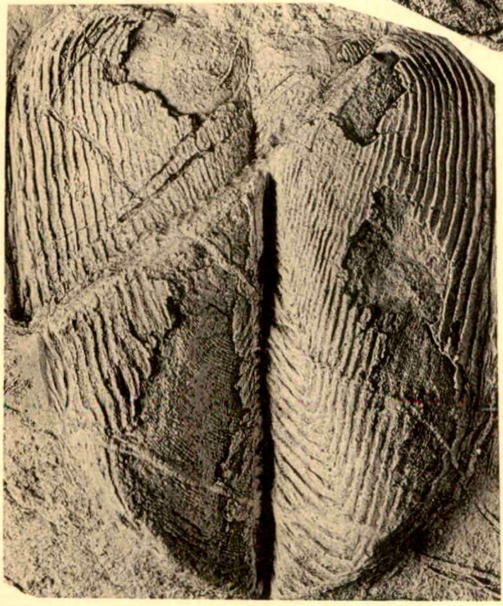
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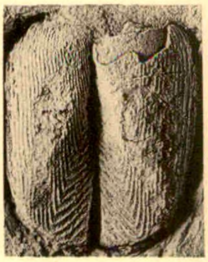
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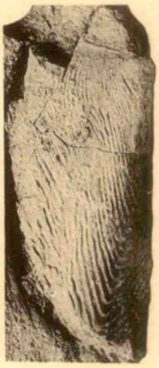
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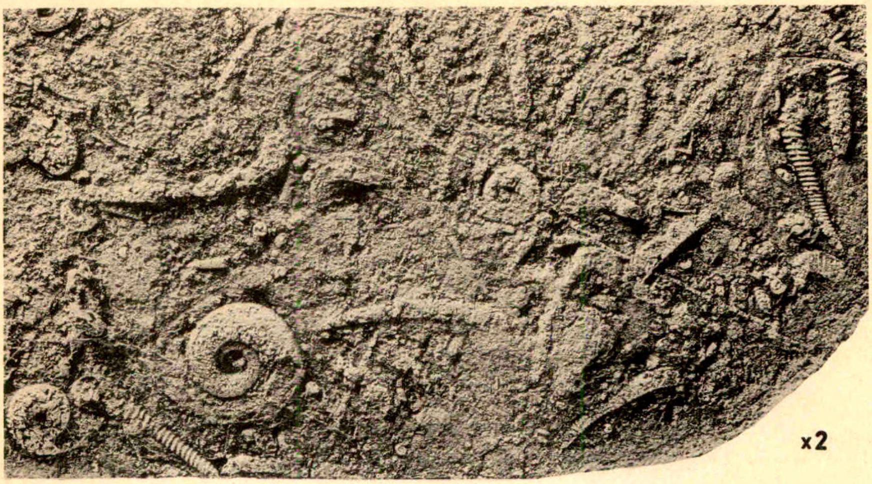
LATE JURASSIC FOSSILS FROM CUBA



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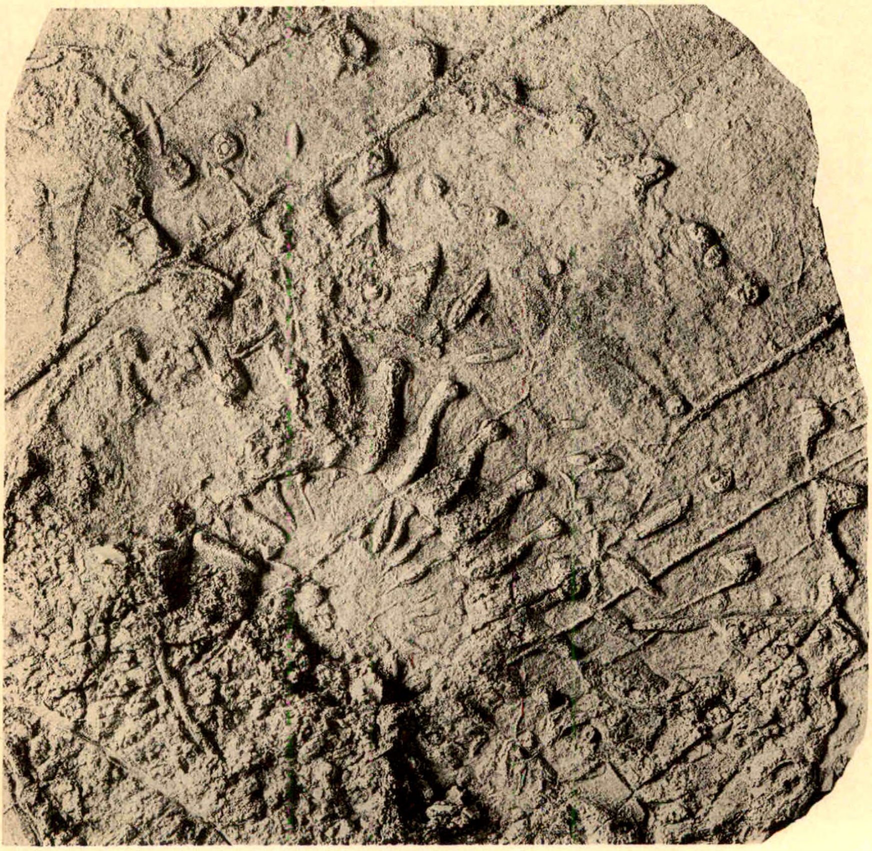

2

LATE JURASSIC FOSSILS FROM CUBA

PLATE 12.—LATE JURASSIC FOSSILS FROM CUBA.

Figure

(1) *Hamulina? rosariensis* Imlay, n sp.

Paratypes on slab U. S. Nat. Mus. 103434f (× 2) in association with *Pseudolissoceras* and *Virgatosphinctes* (opposite side of slab) from Atlantic Refining Company of Cuba locality 5229 (p. 1457).

(2) *Corongoceras* cf. *C. filicostatum* Imlay.

Specimen U. S. Nat. Mus. 103400 from Atlantic Refining Company of Cuba locality 5229 (p. 1449)

Figures natural size unless otherwise indicated.

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NEBRASKAN-KANSAN DRIFT BOUNDARY IN MISSOURI

BY CHAUNCEY D. HOLMES

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ABSTRACT

Development of late-mature topography has removed nearly all Kansan drift from extensive areas in northwestern Missouri, and the present land surface bevels the Nebraskan gumbotil horizon. In north-central Missouri, the Grand River and its larger tributaries have cut through the Kansan deposits, developing a pronounced digitate boundary between the two drifts and leaving numerous Kansan outliers on the higher divides. Kansan gumbotil caps extensive tabular divides east of the Chariton River valley, but many of the deeper intervening valleys expose inliers of Nebraskan drift. In places, the bedrock surface lies higher than the adjacent Nebraskan gumbotil horizon.

Kansan drift apparently overlaps the Nebraskan to the southeast, though present data are inadequate for accurate age determination along much of the glacial boundary.

At some localities the Aftonian horizon is marked by sandy and gravelly silts presumably of fluvial origin. Similar materials of widespread occurrence beneath the Kansan drift rest on bedrock and may be preglacial.

INTRODUCTION

INTRODUCTORY STATEMENT

Mapping of the Nebraskan-Kansan drift contact in northern Missouri was done during the summer of 1941. The principal results have been made available to the National Research Council Committee for a new glacial map of North America. The chief purpose has been to ascertain the present distribution of the two drift sheets, though no attempt was made to verify the precise southern limit of glaciation.

ACKNOWLEDGMENTS

Field expenses were defrayed by a grant from the Penrose Bequest of The Geological Society of America. Before starting the field work, the writer accompanied Dean George F. Kay on a field conference through southern Iowa to observe the Pleistocene deposits so well interpreted in that area. The writer is grateful to Dean Kay for his suggestions and kindly interest in the project.

REGIONAL SETTING

Northern Missouri lies along the periphery of the glaciated area of North America, and its glacial history is closely related to that of Iowa, Kansas, and Nebraska. The Nebraskan and Kansan drift sheets, well known in Iowa, extend into northern Missouri, and their drift-border phenomena are continuous westward from Missouri into Kansas. In contrast, the Mississippi River on the east marks an important Pleistocene boundary (Fig. 1).

TOPOGRAPHY AND DRAINAGE

Approximately the western three-fifths of northern Missouri is drained by streams that rise in southern Iowa and flow generally south and southeast to the Missouri. These include (west to east) the Nodaway, Platte, Grand, and Chariton, with their numerous tributaries. Flood plains 2 miles wide, bordered by low terraces, are common, and the intervening uplands have a late-mature aspect, though a few limited remnants of tabular divides still remain, capped by Kansan gumbotil. Eastward from the extreme north-central part, such tabular divides are more extensive. East of the Chariton River, broad till-plain uplands separate the relatively short tributaries of the Mississippi. Eastward from the junction of the Chariton and Missouri, a strip of territory up to 30 miles wide drains south to the Missouri, and here tabular divides are less common largely because this area lies at, or beyond, the limits of glaciation.

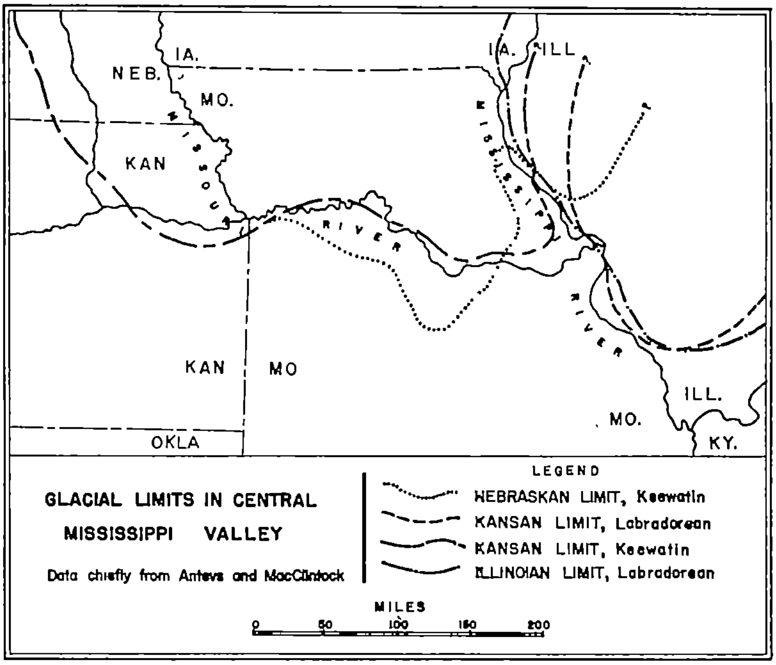
IA.
NEB.
MO.
KAN
MISSOURI
RIVER
IA.
ILL.
MISSISSIPPI
RIVER
KAN
MO
OKLA
MO.
ILL.
KY.
GLACIAL LIMITS IN CENTRAL MISSISSIPPI VALLEY
Data chiefly from Antevs and MacClintock
LEGEND
NEBRASKAN LIMIT, Keewatin
KANSAN LIMIT, Labradorean
KANSAN LIMIT, Keewatin
ILLINOIAN LIMIT, Labradorean
MILES
0
50
100
150
200

FIGURE 1.—*Map of central Mississippi valley region*

Showing approximate limits of Keewatin and Labradorean drifts Adapted from Antevs (1929) and MacClintock (1933)

STRUCTURE AND BEDROCK

Pennsylvanian shales, sandstones, and thin limestones constitute the bedrock over most of the glaciated portion of the State, though Mississippian limestones underlie the eastern and southern parts. The gentle northwest regional dip is varied locally by minor anticlinal flexures which have not appreciably affected glaciation.

PREVIOUS WORK

The first comprehensive report on the Pleistocene deposits of Missouri was published by Todd (1896). Chamberlin (1895) had recognized two glacial stages in Iowa, which he named the Kansan (older) and the Iowan, separated by the Aftonian interval. In the present classification, these are the Nebraskan and Kansan, respectively. Todd found no evidence of multiple glaciation in Missouri, but he regarded Chamberlin's classification as probably valid for northern Missouri as well as for Iowa.

In connection with mapping glacial deposits in southern Iowa in 1920, Kay (personal communication) found both Nebraskan and Kansan drift at a few localities in northern Missouri. In 1924, Shipton described sections of the two drifts in northwestern Missouri, but he included no areal map. Antevs' map (1929, p. 645) shows the southern limits of the two drifts but does not show their present distribution.

In an unpublished work, Trowbridge (1938) concluded that all the drift in northern Missouri is referable to a single glaciation, probably the Kansan.

MAPPING THE DRIFT CONTACT

CRITERIA FOR DETERMINING BOUNDARY LOCATION

General statement.—The Nebraskan gumbotil and its related phases of less-weathered drift constitute the only reliable basis known at present for distinguishing the two drift sheets. Gumbotil development implies a till-plain surface of low relief, and favorably located exposures available at the time of mapping suffice to show the general configuration of the Nebraskan gumbotil surface. An aneriod altimeter was used in areas for which adequate topographic maps were not available.

Nebraskan gumbotil.—The Nebraskan gumbotil was already a discontinuous sheet when the present erosion cycle began. Streams probably destroyed a part during late Aftonian time, and a few masses of Nebraskan gumbotil observed imbedded in Kansan till, indicate erosion by the Kansan glacier. Normally the Nebraskan gumbotil grades downward into a zone of oxidized and leached till generally 1 to 2 feet thick, which in turn grades into oxidized but unleached till. This relation contrasts with the abrupt basal contact of gumbotil masses transported by a later glacier and imbedded in its drift. Nebraskan gumbotil in Missouri ranges from 2 to 15 feet thick, which may or may not approximate the original thickness.

Gumbotil is slightly more resistant to erosion than unleached oxidized till and tends to form benches or breaks-in-slope. The unoxidized and unleached till is likewise relatively resistant but is seldom exposed along gentle slopes such as those on which gumbotil benches occur.

Gumbo silt.—Thoroughly weathered silt with associated sand and sparse gravel apparently underlies the Kansan till in many areas, particularly near the larger river valleys. Many deposits of this type cannot be distinguished satisfactorily from alluvial terrace materials of the present erosion cycle, and no exposure of unleached Kansan till resting on such deposits was observed. Typical exposures occur in the maturely dissected areas bordering the Chariton River valley where orange-colored gumbo sand ranges from 3 to 50 feet thick. In some localities such deposits rest on bedrock at higher-than-average altitude, but in most outcrops the base is not exposed. For example, three outcrops of gumbo silt and sand occur along the graded road in NW. ¼ sec. 18, T. 61 N., R. 15 E., Adair County, Missouri. A limited variety of much-weathered igneous pebbles in the associated gravel lenses indicates derivation from drift. Nebraskan gumbotil crops out in the road immediately upslope from the silt and sand. At another locality, SW. ¼ sec. 23, T. 66 N., R. 14 W., Schuyler County, Missouri, a road cut exposes Nebraskan gumbotil with abrupt lateral gradation into sandy gumbo silt. Weathering of the silt and till was probably contemporaneous, and the contact appears to have been originally the side of a silt-filled channel in the Nebraskan till plain. However, at many other localities the top of the silt and sand deposits lies at, or slightly below, the calculated local altitude of the Aftonian horizon, with known Kansan drift exposed near by at higher altitudes, but with no outcrop of Nebraskan gumbotil in the immediate vicinity. Also, many gravelly deposits lack identifiable erratics, and the genetic relations are therefore uncertain; some may be preglacial.

Secondary carbonate.—Below the Kansan gumbotil in many localities, both the Kansan and Nebraskan drift sheets are traversed by an irregular system of joints filled with chalklike calcium carbonate. The veins are commonly less than half an inch wide as seen in newly excavated road cuts 15 to 30 feet below the surface in mature topography. In general, the veins narrow with depth. Nearer the surface the carbonate forms irregular nodules rather than continuous veins. In many areas solution of these nodules has little more than kept pace with surface erosion. In regions of recent dissection but of slight relief, as in northern Chariton and southern Linn counties, Missouri, the present upper limit of carbonate nodules is less than 10 feet below the base of the Kansan gumbotil. These carbonate

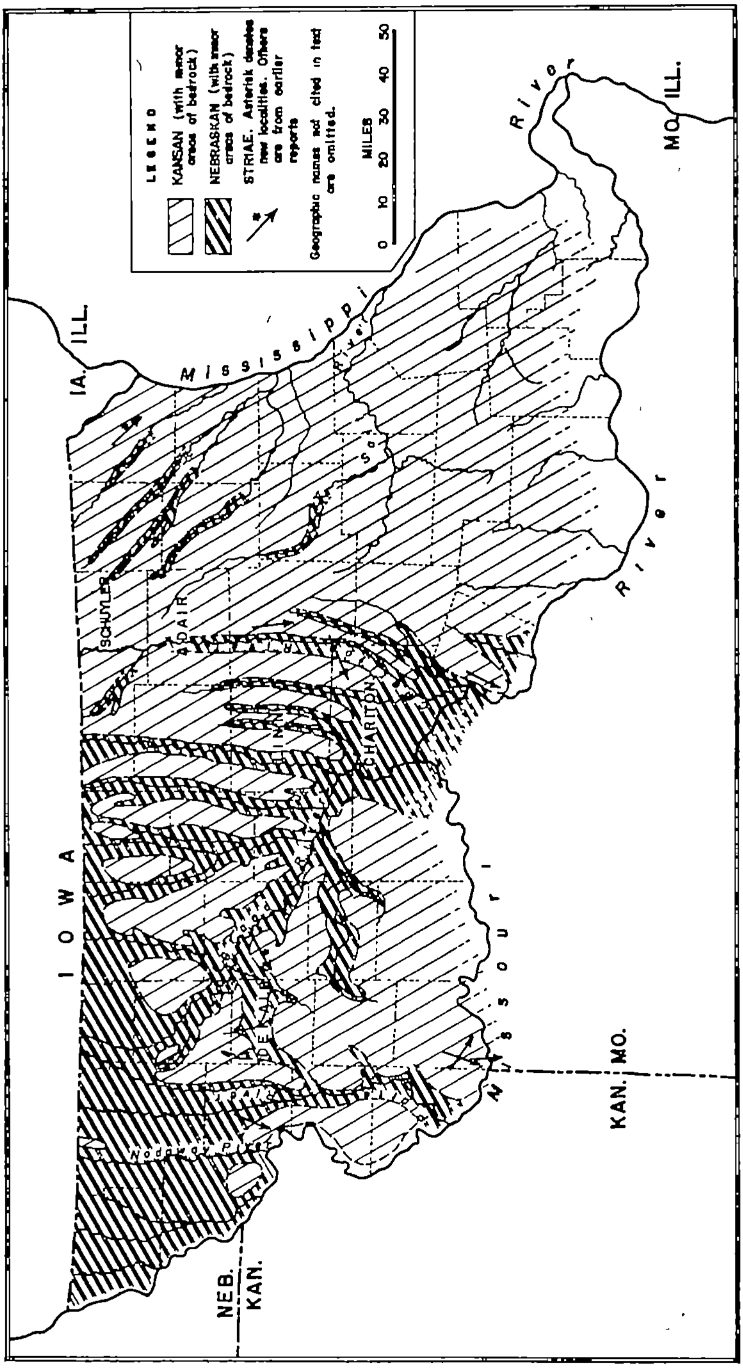
LEGEND
KANSAN (with minor areas of bedrock)
NEBRASKAN (with minor areas of bedrock)
STRIAE. Asterisk denotes new localities. Others are from earlier reports
Geographic names not cited in text are omitted.
MILES
0 10 20 30 40 50
I O W A
IA.
ILL.
MO.
ILL.
NEB.
KAN.
KAN.
MO.
Mississippi
River
Missouri
River
SCHUYLER
ADAIR
CHARITON
DE KALB
Nodaway R.

FIGURE 2.—*Preliminary map of Nebraskan-Kansan drift boundary in Missouri*

masses are as abundant in the Nebraskan gumbotil and in the leached zone beneath as they are in till similarly situated but not previously leached. Some of the secondary carbonate in the Nebraskan till may be pre-Kansan, but most of it is definitely post-Kansan and antedates the present erosional topography.

Secondary carbonate nodules in a gumbotil matrix indicate a previous cover of calcareous material. No such occurrence has been noted in Kansan gumbotil, and no known occurrence was referable to the loess mantle as a source. Hence the presence of secondary carbonate in a gumbotil or in the subjacent leached zone marks the exhumed Aftonian horizon. Evidence of this type is most valuable in mapping maturely dissected regions.

Topography.—In the southeastern part of the region the Kansan gumbotil plain extends into areas where only a single till sheet rests on relatively high bedrock, thus indicating that the Kansan drift overlaps the Nebraskan. In most other places near the drift border, correlation is uncertain either because erosion has destroyed the original surface or because the original surface was somewhat undulating.

MAP OF THE DRIFT BOUNDARY

Figure 2 shows the boundary between the Nebraskan and Kansan drift sheets. In drawing the contact between known outcrops of Nebraskan gumbotil, no allowance has been made for possible valley fills of Kansan drift below the Nebraskan gumbotil horizon. Also, the boundary omits the re-entrants into the smaller valleys.

In the western part of the region the general southward slope of the Aftonian horizon exceeds slightly the corresponding general southward descent of the mature land surface. Practically all the Kansan drift has been removed from the area near the Iowa boundary, and the Nebraskan gumbotil occurs on some hilltops as far south as central DeKalb County, Missouri. Still farther south it crops out lower on the hill slopes. Small outliers of Kansan till probably cap many divides in areas shown as Nebraskan drift.

Kansan drift still covers most of the eastern part of the region, with limited inliers of Nebraskan drift along some of the valleys. These inliers apparently terminate eastward against bedrock.

THE DRIFT SHEETS

GENERAL STATEMENT

The Kansan and Nebraskan drift sheets have been carefully studied in Iowa (Kay and Apfel, 1929). Ice appears to have moved generally

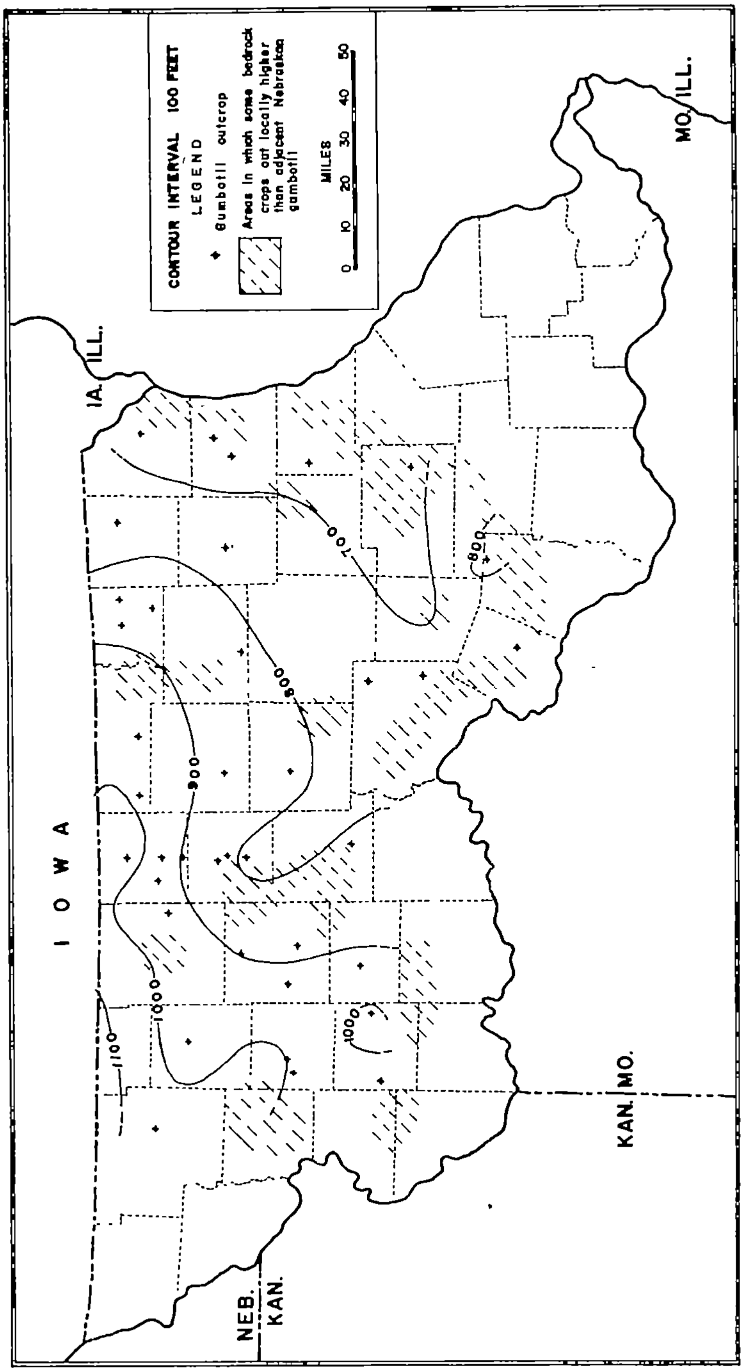
CONTOUR INTERVAL 100 FEET
LEGEND
Gumbotil outcrop
Areas in which some bedrock crops out locally higher than adjacent Nebraskan gumbotil
MILES
0
10
20
30
40
50
IOWA
IA.
ILL.
MO.
ILL.
KAN.
MO.
NEB.
KAN.
700
800
800
900
1000
1000
1100

FIGURE 3.—*Generalized contour map of Nebraskan gumbotil horizon in Missouri*

southward, but striae show some local variation. Most striae occur at the locally higher bedrock altitudes and probably were made by Kansan ice, though definite proof is lacking.

NEBRASKAN DRIFT

Preglacial surface.—The preglacial bedrock relief in northwestern Missouri exceeded 400 feet (Trowbridge, 1938) though the local relief was less in most counties. The major drainage line followed approximately the present course of the Grand River. Few data are available regarding the bedrock surface in northeastern Missouri, but observed areas of outcrops suggest that the preglacial relief was about the same as that farther west.

Thickness.—Nebraskan drift filled the preglacial valleys, though it only mantled some of the higher bedrock hilltops. Thus along the upper part of Chariton River valley the bedrock crops out at altitudes slightly higher than the adjacent Nebraskan gumbotil horizon and is overlain only by Kansan drift. Other similar areas, especially in the east-central and southwestern parts of the region, probably were once mantled with Nebraskan drift. The present maximum thickness of Nebraskan drift may exceed 200 feet locally in northwest and north-central Missouri, but it averages about 50 feet (estimated).

Aftonian surface.—Figure 3 shows the generalized configuration of the Nebraskan gumbotil surface. The main drainage line seems to have been somewhat north of the present Grand River valley and may have continued generally eastward to the Mississippi. The Aftonian valleys were probably shallow.

KANSAN DRIFT

Thickness.—In a few places the Kansan and Nebraskan gumbotils occur within a mile of each other, and the approximate thickness of the Kansan drift can be ascertained. In the northeastern part of the region it is 40 to 50 feet thick, 80 to 100 feet in the north-central part, and 50 to 60 feet in the west-central part. It is much thinner near the drift margin. In many localities the Aftonian horizon is either concealed or has been destroyed. In the latter case the actual thickness of Kansan drift may exceed the computed thickness as measured vertically between the projected gumbotil horizons. The present erosion cycle has removed the Kansan drift from extensive areas in the northwestern part of the State.

Kansan till plain.—Many of the stream courses consequent on the Kansan till plain are nearly straight and subparallel throughout their lengths of more than 50 miles. Their relatively close spacing suggests that the till-plain surface was characterized by shallow, large-scale flutings which may indicate the general directions of flow of the Kansan ice. In part, they may reflect the Aftonian surface and hence be related to the directions of movement of the Nebraskan ice. Valley deepening was probably negligible during Yarmouth time, for the common occurrence of secondary carbonate nodules within 10 feet of the base of the gumbotil implies a fairly high water table and practically stagnant ground water.

PROBLEMATICAL AGE OF BORDER DRIFT

The drift border lies mostly in the dissected area adjoining the Missouri River valley, and many of the isolated patches of till are completely oxidized and leached. The leached zone beneath the Kansan gumbotil averages at least 10 feet, while that beneath the Nebraskan gumbotil averages less than 5 feet. Although the border-zone drift apparently weathered under drainage conditions that resulted in deep leaching and lack of typical gumbotil characteristics, much of the leaching may be post-Aftonian. Therefore Kansan drift may appear as old as Nebraskan. Criteria for determining satisfactorily the age of drift in much of the border zone are lacking.

CONCLUDING STATEMENT

Maximum accuracy in correlating drift exposures requires data from critical localities when favorable exposures become available. The natural deterioration of road-cut exposures is now augmented by landscaping of the excavations. Therefore such outcrops may be studied best while excavation is in progress. A consistent program of study is needed which should yield much additional information about the glacial deposits in Missouri.

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204 Swallow Hall, University of Missouri, Columbia, Missouri.
Manuscript received by the Secretary of the Society, March 11, 1942.
Project Grant 385/41.

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STRUCTURE OF THE UPPER LARAMIE RIVER VALLEY, COLORADO-WYOMING

BY R. H. BECKWITH

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ABSTRACT

The upper Laramie River Valley is at the junction of the Medicine Bow Range and the Colorado Front Range. Pre-Cambrian crystallines and 8000 feet of predominantly shaly Paleozoic and Mesozoic beds are sharply folded and faulted. The valley, a structural trough trending north to northwest, is bounded on both sides for the most of its length by outward-dipping thrust faults. Near the south end, where the trough is deepest, its width is less than the respective slips of the bounding faults. Erosion must have cut away the front of one or both of the advancing thrust blocks during the interval of deformation in order to permit them to reach their present positions.

The following unusual features are described and discussed in terms of genesis: (1) A northeast-dipping thrust carries an anticline onto the adjacent anticline to the southwest and produces thrusting of younger rocks onto older. In the fault zone are slice blocks younger than both formations bounding the fault zone (2) East of this thrust is a north-striking high-angle fault with downthrow on the east. The fault terminates against the thrust to the south and dies out in a monocline to the north. (3) A high-angle fault shows several reversals of relative vertical movements of blocks. (4) Several high- and low-angle faults turn in depth toward stratification and do not affect the crystallines (5) Pre-Cambrian and younger rocks occur in relations that permit deductions on the influence of pre-Cambrian lithology on Laramide structures.

Application of criteria to distinguish underthrust from overthrust faults leads to inconsistent results. It is concluded that features in the mapped area of a kind that have been advanced as criteria show only differential relative movements produced by east-west compression acting on a vertically and horizontally nonuniform crust.

INTRODUCTION

Studies of Rocky Mountain structure by the Geological Survey of Wyoming have covered a large part of the eastern border of the Medicine Bow Range. The detailed mapping, including an area terminating 3 miles north of the Colorado line, was done by the writer. Reconnaissance showed that structural features to the south in Colorado have no counterpart in the border of the Medicine Bow Range in Wyoming and are uncommon in the Rocky Mountains. Exposures of pre-Cambrian rocks in close association with folds and faults affecting Paleozoic and Mesozoic sediments provide more information on pre-Cambrian control of Laramide structure. The present study of an area of about 165 square miles was undertaken as a continuation of previous work.

The geology of the area (Fig. 1) was first mapped in generalized form by the Fortieth Parallel Survey (King, 1876, Map I). The parts of the geologic maps of Colorado (Burbank, 1935) and Wyoming (Campbell, 1925) covering the upper Laramie River Valley are based on reconnaissance and lack detail. The geologic map of the Laramie quadrangle (Darton *et al.*, 1910), a 30-minute quadrangle north of the State line and east of the 106th meridian, is on a scale large enough to show structure fairly accurately. Because of a mistake in stratigraphy and omission of a fault south of Ring Mountain in the geologic map of the Laramie quadrangle, the writer remapped the part of the quadrangle indicated in Figure 1.

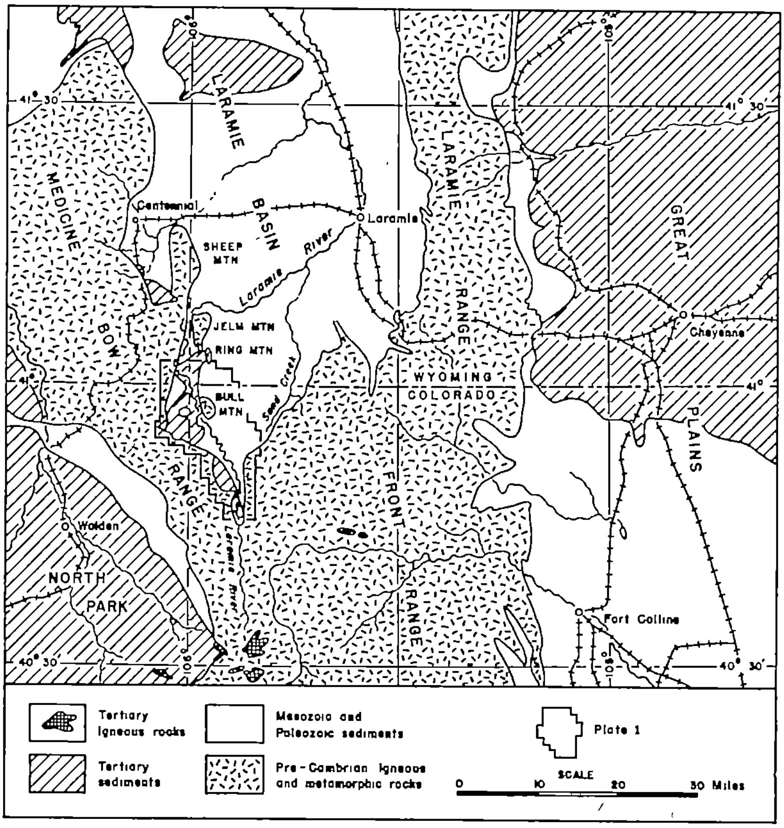

FIGURE 1.—*Index map*

Topographic maps on a scale of 1 : 125,000 cover the upper Laramie River Valley except the part south of the State line and west of the 106th meridian. The contour interval of the Laramie quadrangle is 50 feet; that of the Medicine Bow quadrangle west of the Laramie and the Home quadrangle south of it is 100 feet.

ACKNOWLEDGMENTS

The field work during the months of June to September 1941 was made possible by a grant from the Penrose Bequest of The Geological Society of America, which the writer gratefully acknowledges. Dr. H. D. Thomas, State Geologist of Wyoming, criticized the section on stratigraphy, and Dr. A. F. Hagner of the University of Wyoming aided in the revision of the manuscript. The writer also expresses his appreciation to his field assistants Thomas L. Kirby and Victor B. Gras.

METHOD OF GEOLOGIC MAPPING

Except for a strip 1 to 2 miles wide along the northern border the area (Pl. 1) is covered by aerial photographs taken for the United States Forest Service. North-south overlap of adjacent prints in the line of flight is about 60 per cent; lateral overlap of strips varies from 10 to 80 per cent. The scale is about 3 inches = 1 mile but varies as much as 10 per cent because of ground relief and, in some cases, variation in flight elevation. Some prints are as much as 10° out of orientation.

The geologic map was originally drawn on the prints. Exact orientation of a print was not known, and consequently strikes were plotted roughly and later corrected. Inasmuch as topographic relief was very useful in tracing out boundaries of geologic units, particularly through forest, drawing in the field was done under a stereoscope.

Plane-table ground control was carried out on a scale of 1 inch = 2000 feet, which is about 15 per cent smaller than the average scale of the photographs. The base line in Ts. 9 and 10 N. was laid out along the Laramie River by stadia and checked by rays to points several miles on both sides. The base line for the northern part was chained from the road intersection on the south boundary of sec. 34, T. 11 N., R. 76 W., to the south face of Red Mountain (Pl. 1). Triangulation carried around the east and west sides of the mountain closed on a flag at the northeast end with an error of 75 feet.

Section, quarter, and correction corners shown on the map were located directly on the plane-table sheets. The part of the map outside the area covered by photographs was made by the usual plane-table and stadia methods. The remainder was redrawn on the plane-table sheets from the photographs by precision pantograph. Adjustment of the pantograph as required by variations in scale of parts of a print reduced distortion error to less than 100 feet for most of the area.

In the southwestern part of T. 10 N., R. 76 W., and in a strip about 1 mile by 3 miles along the upper part of Shell Creek in T. 11 N., Rs. 75 and 76 W., the timber is so dense that the plane table could not be used. For these areas prints were fitted together, and the map was pantographed on the sheets using triangulation points several miles apart without adequate check points for distortion. Map locations in these areas are probably much less accurate than in the remainder of the map.

TOPOGRAPHY AND DRAINAGE

The northern part of the upper Laramie River Valley (Fig. 1) is separated from the Laramie Basin by four mountains, Jelm, Ring, Red, and Bull, which form a watershed 1500 to 2000 feet above the river. The lowest point on the crest, at 8100 feet elevation, is at Red Mountain Pass between Ring Mountain and Red Mountain (Pl. 1), and the highest, at 10,000 feet, is at the southeast end of Bull Mountain. An unusual feature of the mountains is that pre-Cambrian crystalline rocks do not form the highest parts. The crest of the northwestern part of Bull Mountain, which consists of pre-Cambrian rocks, is at about 8600 feet. Outliers of Dakota sandstone and Benton shale capping Red Mountain and the central and southeastern parts of Bull Mountain stand from 500 to 1400 feet higher.

From the Dakota and Benton outlier of the southeastern part of Bull Mountain a hogback 500 to 700 feet high formed by the Dakota sandstones and conglomerates extends 3 miles south and joins Green Ridge. Jimmy Creek has cut a steep notch in the hogback and a broader valley in the less resistant sediments to the east. Sand Creek Pass at the head of the valley on the outcrop of the Casper sandstone is at 9000 feet elevation. Green Ridge to the south is a part of the Front Range. Its gently rolling surface a few miles east of the Laramie River varies between 10,000 and 10,400 feet. The western edge is a scarp 1500 feet high with slopes up to 30° for hundreds of feet.

The flood plain of the Laramie River rises from about 7800 feet at the north boundary of the map (Pl. 1) to 8350 feet at the south boundary. The northern 2 miles of the stream is in a narrow winding canyon cut in resistant pre-Cambrian gneisses. A few thousand feet to the east horizontal Tertiary beds lie unconformably on pre-Cambrian and younger rocks. The winding course of the canyon and the horizontal cover in the immediate vicinity clearly indicate a superposed stream. Southward the valley floor widens to about 4 miles in the southern part of T. 11 N., where much of it is covered by thin terrace gravels. The present flood plains of the Laramie River and its larger tributaries are 50 to 100 feet below the terrace surface and locally are several thousand feet wide. Northeast from the floor the land surface rises abruptly to Bull Mountain. To the west and southwest the surface carved principally on Tertiary sediments rises gradually and, in some places where the Tertiary overlaps onto the pre-Cambrian, merges with the gently rolling surface of the Medicine Bow Range at elevations of 8500 to 9000 feet.

Between the Front Range and the Medicine Bow Range in the southern part of the area are two longitudinal valleys and an intervening ridge. The northern part of the ridge, North Middle Mountain, has a flat top at about 8600 feet elevation. The north and south ends carved from pre-Cambrian rocks project several hundred feet higher. Middle Mountain is a rounded ridge that terminates in a sharp peak of pre-Cambrian rocks at 9300 feet. The northern part of the western valley is drained by McIntyre Creek, and the southern part by Stub Creek and the Laramie River. The eastern valley is occupied by the Laramie River, Nunn Creek, and Porter Creek, a short stream that flows south along the east side of Middle Mountain and joins the Laramie River 2 miles south of the map boundary. The writer does not attempt to explain the peculiar drainage pattern but suggests that it may be genetically connected with superposition from Tertiary cover and glacial damming, because the single valley south of Middle Mountain is a glacial trough.

PRE-CAMBRIAN ROCKS

The pre-Cambrian rocks were not mapped in detail. The aerial photographs, however, yield considerable information on the character and structural features of the pre-Cambrian rocks; where these seemed to have influenced Laramide structures detailed observations were made.

Among the oldest rocks are quartz-biotite schists. In a few places muscovite is fairly abundant, indicating that the rocks originally may have been sediments. The schists are cut by accordant and transgressive tabular masses of metadiabase showing, in many places, well-developed parallel alignment of feldspar and hornblende crystals. Metamorphics of undoubted sedimentary origin, such as quartzites and marbles, were not observed in place or as pebbles in gravels carried by streams draining the more remote parts of the mountains. Apparently the thick quartzites and marbles in the central part of the Medicine Bow Range west of Centennial (Blackwelder, 1926) do not extend into the southern part. The quartz-mica schists are probably, in part at least, the equivalent of the Idaho Springs formation of the central Front Range (Lovering, 1935, p. 6-10). Lovering showed that the formation was originally sedimentary and that schistosity parallels stratification. He states (1929,

p. 64) that the thickness of the sediments has been estimated by Clarence King as not less than 25,000 feet and that no more accurate figure is available.

Hornblende schists and greenstones are also found in the pre-Cambrian complex and are cut by metadiabases and irregular masses of basic rock coarse enough to be classed as gabbro. These rocks represent a period of basic extrusive and intrusive igneous activity during which the metadiabases cutting the mica schists were intruded. The basic rocks are probably at least partial equivalents of the Swandyke hornblende gneiss of the central Front Range (Lovering, 1935, p. 10-11) and the hornblende schists (Spock, 1928, p. 185) in the northern part several miles south of the head of the Laramie River.

The mica schists and basic igneous rocks have been invaded by a granite batholith. Locally massive coarse pink granite is exposed over areas of several square miles,—apparently representing cupolas. Other exposures of granite show numerous schist and metadiabase xenoliths, some of which are only slightly altered and others so completely digested that they are merely patches of granite containing abnormally high percentages of ferromagnesian minerals. In many places the schists are cut by aplite and pegmatite dikes that interfinger with the schists, showing an early stage of injection. The most extensive areas of granitic rocks consist of pink and gray gneisses formed by *lit-par-lit* injection of the schists. Within the gneisses, even where the original schists have been almost entirely absorbed, tabular and irregular masses of metadiabase and other basic intrusives at first appear to be dikes. Actually they are cut by granite dikelets and consequently are older than the granites; the tabular basic intrusives are remnants that resisted injection because of their poor cleavage.

STRATIGRAPHY

GENERAL STATEMENT

The stratigraphic section (Table 1) includes approximately 8200 feet of Pennsylvanian to Upper Cretaceous beds. Disconformities in the succession produce no noticeable discordance. Two angular unconformities occur, one at the base of the Fountain, and the other at the base of the Tertiary. The Tertiary laps onto all older rocks.

Thickness of the Fountain was obtained from the width of outcrop where the beds are nearly vertical in the western part of T. 12 N., R. 77 W. (Pl. 1). The section from Casper to Morrison was measured by Brunton and steel tape in secs. 8 and 9, T. 11 N., R. 76 W. Because of poor exposures, sections of the Dakota and Benton groups were not measured; the figures given are for a locality 18 miles north of the State

TABLE 1.—*Geologic formations of the upper Laramie River Valley*

Age	Unit mapped	Formation	Thickness (feet)	Characteristics
Quaternary	Quaternary Alluvium		0–100(?)	Floodplain and pediment gravels, landslip and glacial debris, and probably some thin remnants of Tertiary sediments.
Miocene(?) and Oligocene(?)	Undifferentiated Tertiary		0–400(?)	While volcanic ash beds, and conglomerates and talus breccias with an ashy matrix. Continental.
	Unconformity			
Upper Cretaceous	Mesaverde formation	Mesaverde formation	2000 +	Brown, gray, and white sandstones alternating with gray sandy shales; contains "ironstone" concretions up to 10 feet across. Marine.
	Steele shale	Steele shale	2400	Dark-gray shales and sandy shales. A zone of sandstones 150 feet thick, 500 feet below the top, is mapped as the Shannon sandstone member. Marine.
	Niobrara formation	Niobrara formation	340	Buff shaly limestones at the base, near the middle, and at the top. Above and below the middle limestone are gray to black calcareous shales. Marine.
	Benton group	Carlile shale	250	Black and gray shale grading upward into calcareous shale. Marine.
		Frontier formation	650	Black shale. At top is gray to brown Wall Creek sandstone about 10 feet thick. Black septarian concretions up to 2 feet across in the lower part. Marine
		Mowry shale	120	Black siliceous shale; weathers silver gray. Bentonite seams up to a foot thick. Marine.
		Thermopolis shale	40	Black shale. Marine.
Lower Cretaceous	Dakota group		350	Upper gray quartzitic sandstone; 30 feet. Upper-middle black shale and shaly sandstone; 150 feet. Lower-middle flaggy brown sandstone; 100 feet. Lower white conglomeratic sandstone and gray conglomerate with rounded chert pebbles up to half an inch across; 70 feet Forms a ridge. Partly marine.
Upper Jurassic	Morrison formation	Morrison formation	270	Shales of various colors containing thin sandstone and limestone lenses, and bands of chert concretions Continental.
	Jelm formation	Jelm formation	245	Buff sandstones; 65 feet. Maroon and white cross-bedded sandstones and maroon conglomerates and shales; 40 feet. Maroon shale; 20 feet Massive orange sandstones; 120 feet. Continental.
Triassic	Chugwater formation	Chugwater formation	575	Brick-red sandy shales and shaly sandstones. Several gypsum beds in the lower part. Continental.
Permian	Forelle limestone	Forelle limestone	10–20	Upper gray to dove crenulated limestone; middle red sandy shale; lower limestone similar to the upper. Marine.
	Satanka shale	Satanka shale	95–133	Red sandy shale. Beneath it around Red Mountain is a 50-foot gypsum bed. In many places a gray to pink limestone up to 3 feet thick occurs at the base. Partly marine.
Pennsylvanian	Casper sandstone	Casper sandstone	110	Buff festoon cross-laminated sandstone grading downward into buff flaggy sandstone. Forms a ridge. Marine (?).
	Fountain formation	Fountain formation	700	Upper 100 feet of flaggy red and white sandstones and red sandy shales; lower part of red shales, sandstones, arkosic grits, and conglomerates. Continental.
	Unconformity			
Pre-Cambrian	Pre-Cambrian			Pink massive granite, gray to pink injection gneisses, black and green hornblende and mica schists, and metadiabases.

line in sec. 19, T. 15 N., R. 77 W. (Beckwith, 1938, p. 1519). The Niobrara was measured by Brunton and steel tape in the NW ¼ sec. 17, T. 11 N., R. 76 W. Thickness of the Steele was obtained from scaled widths of outcrops and dips in secs. 2 and 11, T. 10 N., R. 76 W. The minimum thickness for the Mesaverde was estimated from dips and the distance in secs. 23 and 24, T. 10 N., R. 76 W., from beds in the lower part of the formation to the inferred position of the axis of the Laramie River syncline. The larger figure for thickness of the Tertiary is based on topographic relief of the area dissected by Forester Creek and Grace Creek.

NOTES ON NOMENCLATURE

The stratigraphy is well known. Many of the units of Table 1 are identical with those of Darton *et al.* (1910, p. 5-11). The stratigraphic subdivisions previously used by the writer (1938, p. 1519-1522) in the adjacent area to the north required little revision. The following notes correlate terminology with that used in surrounding territory.

The Casper formation of Darton *et al.* (1910, p. 5-6) was divided by Knight (1929) into Fountain and Casper. The writer (1938, p. 1521), using the term Casper as restricted by Knight, recognized the following members:

Member	Thickness (feet)
Buff festoon cross-laminated sandstone	80
Gray sandy shale	30
Red and white shaly sandstone	100
White festoon cross-laminated sandstone	70
Total	280

The upper sandstone extends to the southernmost exposures in T 9 N. (Pl. 1). The gray sandy shale becomes more sandy southward and is apparently represented by the flaggy sandstones in the lower part of the measured section (Table 1). The term Casper, as here used, consequently includes only the upper two members of the Casper in the area to the north. The red and white shaly sandstones are present in the upper Laramie River Valley but were included in the Fountain because of the difficulty of distinguishing them from the underlying beds. The lower white sandstone extends into the northern 2 miles of T. 12 N. but was mapped as part of the Fountain. This makes little difference in adjoining maps west of Ring Mountain, where the Casper dips steeply. On the east side, however, where dips are low, the base of the Casper is shown in the SE ¼ sec. 32, T. 13 N , R. 76 W. (Beckwith, 1938, Pl. 1), but not immediately to the south in sec. 5, T. 12 N., R 76 W. The white sandstone present in this section extends southwest and terminates against Ring Mountain fault in sec. 7.

The term Satanka is used in the same sense as by others. There has been, however, much controversy about identification of the formation around Red Mountain. Darton *et al.* (1910, p. 7) say:

> "The Satanka shale lies between the Forelle limestone and the Casper formation along the west slope of the Laramie Mountains. East of Laramie the shale lies on limestone that forms the top of the Casper formation, but farther north

and south it rests on massive gray sandstone. The shale is absent in the region between Sand Creek and Red Mountain, where the Forelle limestone lies directly on light-colored sandstone that is believed to be at the top of the Casper formation."

Thomas studied the Permian and Lower Triassic red-bed succession extensively from central to southeastern Wyoming. He gives a detailed section (Thomas, 1934, p. 1690) measured at Gypsum Butte, which is shown on the map (Pl. 1) by the Forelle outliers in the NE ¼ sec 9, T. 12 N., R. 76 W. He identifies the 15-foot limestone capping the butte and extending 4 miles along the north base of Red Mountain as Forelle. In the Satanka well exposed beneath it he lists 72 feet of red sandy shale, 53 feet of gypsum, 5 feet of pink shaly limestone, and 3 feet of olive shale resting on the Casper. He says (Thomas, 1934, p. 1689):

> "Misstatements as to the stratigraphy of the Satanka, the Forelle, and the lower part of the Chugwater at Red Mountain have become almost hopelessly entangled in the geologic literature. A fossiliferous limestone, which in places is a breccia, was mistaken by Darton and Siebenthal for the Forelle. Since the limestone occurs only three feet above the top of the Casper, they assumed that the Satanka was absent at Red Mountain."

The writer's observations corroborate those of Thomas. The red sandy shale of the Satanka is present in the upper Laramie River Valley to the southernmost exposures of sediments. The gypsum bed thins out in a short distance from Red Mountain; it is present in the exposures in the W ½ sec. 29, T. 12 N., R. 76 W., and disappears in the 2 miles of outcrop to the east. The lower limestone of the Satanka is locally present at least as far south as the southern part of T. 11 N. The writer's map of the area around Red Mountain is markedly different in some places from that of Darton *et al.* (1910, Areal Geology) because of their failure to recognize the Satanka.

The term Chugwater, as used by Darton *et al.* (1910, p. 7-8), designated all the red beds above the Forelle. Knight (1917) recognized that the upper 250 feet is separated from the lower part by a disconformity and, for the part above the disconformity, proposed the name Jelm formation. The writer found that this unit, consisting of resistant sandstones and conglomerates between the Chugwater red shales and the Morrison variegated shales, is convenient for mapping. Heaton (1939) has shown that the upper part of the Jelm is the Entrada (Jurassic) sandstone and probably the equivalent of the basal sandstone of the Sundance. He says (Heaton, 1939, p. 1158):

> "Section No. 1 at Bull Mountain, Wyoming,[1] shows nearly 300 feet of massive sandstone, the upper half of which is quite different in appearance from the lower half and is separated from it by a layer of limestone and red shale. The entire thickness is undoubtedly Knight's Jelm formation, but it seems quite likely that the Entrada is represented in the upper part."

The Morrison formation is so well defined and generally recognized that no comment is necessary.

The term Cloverly formation was applied by Darton *et al.* (1910, p. 9) to the two lower members of the Dakota group (Table 1); they included the two upper members in the Benton group. Because of the convenience of drawing contacts at the base of a well-defined conglomerate or conglomeratic sandstone and at the top of a ridge-forming sandstone, the writer preferred to place all the sandstones in one stratigraphic unit and to use the term Dakota group as defined by Lee (1927, p. 18).

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[1] Apparently a mistake in location with respect to the State line, as Bull Mountain is in Colorado.

The term Benton group, as here used, is a restriction of the term Benton shale of Darton *et al.* (1910, p. 9), as the top of the Dakota group is drawn higher than the top of the Cloverly formation. Although they recognized several members in the Benton shale, they did not completely subdivide it. The formations of the Benton group in Table 1 are essentially those recognized by Dobbin *et al* (1929b, p. 139) in the Rock Creek oil field east of the Medicine Bow Range about 50 miles north of the State line; their Thermopolis shale, however, also includes the upper two members of the Dakota group. The writer's map (Pl. 1) does not show the formations of the Benton group because they can be distinguished only in areas of excellent exposures; they were recognized in many places and were useful for structural interpretation.

Thomas (1936) has shown that the Carlile shale is of Niobrara age. The writer did not, however, include it in the Niobrara because, in most places, the Wall Creek sandstone immediately below is not well enough exposed for mapping.

No revision is made in the terms Niobrara, Steele, and Mesaverde as defined by Darton *et al.* (1910, p. 9-10).

The Tertiary consists largely of material derived from rocks in the immediate vicinity. In most places the fragments in the conglomerates and breccias are granites and metamorphic rocks. West of Red Mountain the Tertiary contains Dakota sandstone and conglomerate boulders several feet across. At the north end of North Middle Mountain it contains pebbles of extrusive igneous rocks several inches across; these increase in size southward to more than a foot at the south end of the mountain. Specimens were collected from a road cut in the NW ¼ sec. 35, T. 10 N., R. 76 W., where the boulders are undoubtedly embedded in ashy material. Thin sections show that the rocks are rhyolitic tuffs, flow-banded rhyolitic obsidians, and porphyritic latites. Spock (1928, p. 214-235) describes volcanics of these types in place about 20 miles to the south.

The Tertiary is best exposed in the valley of Grace Creek, but a complete section cannot be measured. The lower part seems to consist of fine ash beds; the upper part contains ashes and conglomerates. The lithologic succession suggests that the Tertiary here is the equivalent of the Brule clay of Oligocene age and the Arikaree formation of Miocene age described by Darton *et al.* (1910, p. 10-11) in the area west of Cheyenne (Fig. 1). The same general lithologic succession can be recognized in the southern part of the Centennial Valley. A jaw and teeth found here in the lower beds have been identified as belonging to the genus *Leptomeryx*, showing that the sediments are almost certainly upper Oligocene (Beckwith, 1938, p. 1522).

TIME OF DEFORMATION

Direct evidence in the upper Laramie River Valley shows only that deformation occurred after deposition of the Mesaverde formation and before the Tertiary sediments of probable Oligocene and Miocene age. In the Hanna Basin beyond the north end of the Medicine Bow Range there were, during this interval, two periods of deformation. The coal-bearing Hanna formation of early Eocene age (Dobbin *et al.*, 1929a, p. 25) about 5 miles north of the range contains granite pebbles, showing that pre-Cambrian rocks were exposed to the south at this time. The Hanna formation overlaps thousands of feet of Cretaceous beds, lies on rocks as old as the Dakota, and is folded. In North Park (Fig. 1) the same history is shown. Near Walden the coal-bearing Coalmont formation, now regarded as unquestionably of Eocene age (Wilmarth, 1938, p. 474), contains conglomerates with quartz, limestone, chert, and schist pebbles up to 3 inches in diameter (Beekly, 1915, p. 55); Paleozoic and pre-Cambrian rocks in the southern part of the Medicine Bow Range were consequently exposed during the Eocene. The Coalmont overlaps several thousand feet of Pierre shale, the approximate equivalent of the Steele and Mesaverde (Table 1). It rests on rocks as old as Niobrara, is sharply folded, and is faulted against rocks as old as pre-Cambrian (Beekly, 1915, Pl. 12).

It is inferred that deformation in the vicinity of the junction of the Medicine Bow Range and Front Range began with arching and erosion of the arch while the late Cretaceous sea still persisted at short distances from the present mountains. Folding and thrust faulting occurred during early Eocene and continued long enough to affect beds of this age but ceased before Oligocene time.

DEPTH OF BURIAL

The folded sediments in the upper Laramie River Valley are about 8000 feet thick. Farther north the Mesaverde is succeeded conformably by 3000 feet of marine Lewis shale and several thousand feet of grits, sandstones, carbonaceous shales, and coals constituting the Medicine Bow formation. The Medicine Bow is overlain unconformably by the Hanna formation. At the Citizen's Coal Mine 5 miles north of Sheep Mountain (Fig. 1) the lower beds of the Medicine Bow contain conglomerates with pebbles of Dakota sandstone and Mowry shale several inches across. Thousands of feet of marine beds must therefore have been stripped from the adjacent rising arch by early Medicine Bow time. A similar conclusion is reached for the region to the south. Lovering (1935, p. 50) states:

> "Before the end of Pierre time the central part of the Front Range highland was pushed above the level of the sea, and the recently deposited shales were exposed to erosion. They were reworked into the upper part of the marine Cretaceous, and the Dakota sandstone was also exposed and reworked at many places and was probably the source of much of the sandy material found in the Fox Hills sandstone."

The short distance of 45 miles from the Citizen's Coal Mine to the south end of the area mapped by the writer suggests that at least part of the Lewis was deposited over the area. A probable minimum thickness of sedimentary cover when deformation began is therefore about 10,000 feet; probable maximum is 15,000 feet. Pre-Cambrian pebbles in the Coalmont formation only 20 miles west and folds and faults affecting it strongly indicate complete removal of sedimentary cover from some of the mountains bordering the upper Laramie River Valley during the later stages of deformation.

COMPETENCE OF ROCKS

The structural pattern produced in an area during an orogeny is dependent upon a number of interrelated factors such as confining static pressure, physical properties and positional relations of different rock masses, and association with igneous phenomena. It is proposed to point out facts that seem to explain, in a measure, some of the characteristics of the structural pattern of the area described.

Stratigraphic evidence indicates that at the beginning of deformation the sedimentary cover was 10,000 to 15,000 feet thick. These figures are low in comparison with thicknesses up to 40,000 feet in the geosyncline of western Wyoming and eastern Idaho. Confining static pressures and temperatures were consequently comparatively low in the sediments and the part of the pre-Cambrian now exposed to observation. During the later stages of deformation the sedimentary cover was much thinner, and pre-Cambrian rocks were locally exposed.

Tertiary igneous rocks do not occur in the area studied. Some of those shown in Figure 1 are of the same age as the horizontal Tertiary sediments. The two bodies to the east and one to the south are indicated on the geologic map of Colorado (Burbank, 1935) as early Tertiary intrusives. It is doubtful, however, that they have had any effect on structure of the area of Plate 1, as its features are similar to those found far to the north in the Medicine Bow Range (Beckwith, 1938; 1941a), where Tertiary igneous rocks are absent.

The various types of pre-Cambrian rocks would necessarily deform differently. Although the writer knows of no experimental work on mode of deformation of metamorphic rocks under stress applied in a direction different from that which produced metamorphism, it is probable that schists with excellent cleavage, under some conditions, deform by folding much as do well-stratified sedimentary rocks; gneisses tend less to slip along banding and are less subject to folding without faulting; and massive granites deform mainly by faulting or movement along closely spaced joints. Evidence is given later to indicate that the core of a sharp anticline involving the sediments is a schist and that the block above an adjacent major thrust fault is mainly massive granite. The writer believes that such differences in behavior of the pre-Cambrian complex are responsible for some of the structures from which oil is produced in the Rocky Mountain region (Beckwith, 1941b, p. 2189).

The folded sedimentary rocks are predominantly shaly. Thick massive limestones, such as those in the lower part of the succession in western

Wyoming and eastern Idaho, are conspicuously absent. This fact and the small superincumbent load seem to account, in part at least, for the absence of flat thrusts known in the deeper part of the Rocky Mountain geosyncline to the west (Mansfield, 1927; Pierce, 1941). Crustal shortening east of the Medicine Bow Range is apparently much smaller and was produced by sharp folding, some of which undoubtedly affects the pre-Cambrian rocks, and by movement along high-angle thrust faults. The sedimentary rocks transmitted horizontal compression only locally and were bent and broken largely in response to movement of the adjacent pre-Cambrian rock masses.

Beds of unusually low strength and resistance to slipping along stratification are found in two parts of the succession. In the area around Red Mountain a gypsum bed up to 50 feet thick occurs 800 feet above the pre-Cambrian in the lower part of the Satanka. Gypsum beds several feet thick occur in the lower Chugwater less than 100 feet higher in the succession; they extend through most of the area. Bentonite seams are abundant in the lower part of the Benton group about 2500 feet above the pre-Cambrian. They seem to have acted as a lubricant and played a part in the localization of faults.

The cross-bedded sandstones and conglomerates of the Jelm and Dakota would not slip easily along stratification and have sufficient strength to transmit stress and support part of the weight on the crest of an arch; they are also brittle. Several faults affecting this part of the succession and apparently dying out downward in the lower Chugwater were formed by fracturing of brittle beds and downward propagation of the fault into others that glide easily.

STRUCTURE

ADJACENT AREA TO THE NORTH

The area north of Plate 1 has been mapped and described (Beckwith, 1938). Because of the Tertiary cover west of Ring Mountain, it seems advisable to summarize some of the structural features exposed to the north.

The narrow trough between the Medicine Bow Range and Jelm Mountain (Fig. 1) contains folded Paleozoic and Mesozoic sediments. On the west the trough is bounded by the West Woods fault, a westward-dipping thrust along which the pre-Cambrian moved east onto beds as much as 3500 feet above the base of the succession. The fault either dies out southward or turns west into the pre-Cambrian, for the full thickness of the Fountain is present at the north boundary of Plate 1. The trough is bounded on the east by the East Woods fault, which is believed to be a westward-dipping subsequent shear thrust along which the beds were carried up to the east onto the pre-Cambrian of the west flank of Jelm Mountain. This fault has a displacement of at least 1 mile where it passes beneath the Tertiary at the southwest end of Jelm Mountain. One mile north of Plate 1 a syncline and an anticline occur between the Woods faults. The syncline is undoubtedly the northward continuation of Stuck Creek syncline. The anticline to the east is thrust onto Jelm Mountain anticline.

Jelm Mountain is bounded on the east by a westward-dipping thrust that apparently dies out in the limb of an overturned anticline somewhere west of Ring Mountain. The north end of Ring Mountain is an anticline with no evidence of faulting. Several thousand feet east an eastward-dipping thrust brings to the surface a small mass of pre-Cambrian not shown in Figure 1, to which the name East Ring Mountain was given.

The exposed pre-Cambrian rocks in the belt 2 miles wide north of the junction of the maps seems to explain some of the peculiarities in structural behavior. The eastern edge of the Medicine Bow Range consists mainly of gneiss, a type likely to deform by fracture. In the south end of Jelm Mountain the pre-Cambrian rocks are schists with well-developed cleavage, thus explaining the apparent transition from a thrust fault at the middle of the east side of the mountain to an overturned fold at the southeast end. The two sharp folds in the sediments west of the south end of Jelm Mountain are probably genetically related to extension of the schists in this direction. The pre-Cambrian rocks in the folded north end of Ring Mountain are schists and metadiabases. The rocks of the East Ring Mountain fault block are mainly massive granites.

RING MOUNTAIN AND RED MOUNTAIN PASS AREA

The south end of Ring Mountain (Pl. 1) consists of massive pink granites and gneissoid granites containing undigested remnants of metadiabase. The overlying beds dip in general to the south and outline a southward-plunging anticlinal nose. In and near the pre-Cambrian, Ring Mountain fault is close to the crest of the anticline. From here it strikes southwest diagonally across trend of the fold into the west flank. South of Red Mountain Pass dips in the west flank steepen westward up to 70° in the Niobrara near the center of sec. 13. In both blocks, beds tend to parallel the fault as they approach it. The Forelle in the SW ¼ sec. 6 is affected by local folds trending parallel to the fault. A syncline is shown on the map; east of the syncline are two folds too small to be shown. Crumpling in the immediate vicinity of the fault points to compression normal to fault strike.

The block east of the fault is upthrown. At the base of the Fountain in the east block the stratigraphic separation (Reid *et al.*, p. 174), or thickness of beds faulted out, is about 950 feet. Where the Casper and Dakota are in contact west of the center of sec. 7, stratigraphic separation is about 1300 feet. Net slip of the fault therefore apparently increases southward. Near the southwest corner of sec. 7 Morrison on the east is in contact with middle Dakota black shale; here less than 100 feet of beds is faulted out, and to the south the fault does not offset the top of the Dakota. The rapid southward decrease in stratigraphic separation in the SW ¼ sec. 7 indicates that the fault cutting out part of the Dakota is merely a branch of a fault farther west beneath the Tertiary. An assumption that Ring Mountain fault dies out in the vicinity of the southwest corner of sec. 7 would also fail to account for continued steepening of westward dips in the west flank of the anticline toward the center of sec. 13.

The dip of Ring Mountain fault could not be obtained. Where its surface trace ascends from the broad valley south of Ring Mountain toward Red Mountain Pass the trace swings to the west, indicating dip to the east. That Ring Mountain fault is an eastward-dipping thrust, as shown near the center of section A-A′ (Pl. 1), is indicated by the fact that the east block is upthrown, the evidence of eastward dip of fault, the minor folds indicating compression normal to fault strike, and the increase in dips of beds in the west flank of the anticline as the fault is approached. Ring Mountain fault is probably the northward continuation of Bull Mountain fault, which is undoubtedly an eastward-dipping thrust.

NORTHWESTERN AREA

Along the eastern edge of the Medicine Bow Range beds in the lower part of the succession are vertical or dip at high angles to the east. Within half a mile to 2 miles the beds flatten eastward into Stuck Creek syncline. In the zone of marked flattening minor folds affect the Morrison and Dakota. In the SW ¼ sec. 33, T. 12 N., R. 77 W., an anticline and a syncline account for the large width of outcrop of the Morrison. Two miles south in secs. 9 and 10 the Morrison appears to have a width of outcrop of half a mile; small folds concealed by the Tertiary seem to account for this. In addition a fault, probably a thrust of not more than 200 feet net slip, omits part of the Benton group in the outcrop to the east. Farther south in Grace Creek Canyon the Dakota is crumpled into four folds in a belt half a mile wide. To the southwest is a small thrust fault, possibly a branch of North Middle Mountain fault.

Minor folding of the Morrison and Dakota in the zone of rapid flattening is related to lithology. The Jelm and Dakota sandstones and conglomerates have a higher crushing strength than the adjacent shales, cross-bedding inhibits slipping along stratification, and when the Jelm and Dakota are sharply bent they fracture by tension. The beds below, between, and above them are shales (Table 1). The sandstones and conglomerates in the succession from Jelm to Dakota seem to have been unable to accommodate themselves to the general movement of rocks involved in the folding of the major anticline and syncline. The western parts of sections B-B′ and C-C′ (Pl. 1) showing minor folds are drawn with crestal thickening of shales in the anticlines, and the folds consequently die out downward.

Stuck Creek syncline is a broad open fold. The surface trace of its axial plane is convex to the west and roughly parallel with Bull Mountain and Ring Mountain faults and the east edge of the Medicine Bow Range. Southward from the vicinity of the State line the syncline plunges south to southeast, partly accounting for the southward increase in the distance from the edge of the Medicine Bow Range to the axis of the syncline. The syncline probably continues farther south than shown on the map, at least to the exposure of Shannon sandstone on Forester Creek. Its south end is apparently cut off by North Middle Mountain fault. North of the State line the Chugwater is the youngest formation in the core, and the syncline is nearly horizontal. Beyond the map boundary it plunges north because beds as young as Benton appear in the center.

Tertiary beds completely cover the area from the Laramie River to the west sides of Ring Mountain and Red Mountain. Inferences on structure of the concealed rocks are, however, possible. The Niobrara at the State line dips west toward a syncline. This syncline cannot be Stuck Creek syncline because the Dakota to the west at the mouth of Stuck Creek is older than the Niobrara. East of Stuck Creek syncline, consequently, there must be an anticline and another syncline. The same line of reasoning applied to the area between the Dakota exposure south of Red Mountain Pass and the Chugwater at the mouth of Beaver Creek shows that west of Ring Mountain anticline there must be a syncline and an anticline; these are shown in the western part of section A-A′ (Pl. 1). In view of the complicated structure exposed 3 miles north, which is shown in detail elsewhere (Beckwith, 1938, Pl. 1, sec. A-A′), relations shown beneath the Tertiary are probably of little value except as generalizations.

RED MOUNTAIN AREA

The top of Red Mountain is capped by an outlier of resistant Dakota sandstone and conglomerate. Nearly horizontal beds down to the Forelle are exposed on the north, east, and south sides. The Chugwater red beds here form bare almost vertical cliffs hundreds of feet high. The gentler west slope of the mountain is nearly parallel to stratification in the Dakota. West of the outlier, however, the Dakota has been removed, and the Morrison is exposed in a belt about 1 mile wide. Farther west the Dakota and Benton steepen, and near the center of sec. 13 the Niobrara, excellently exposed in a deep gulch, dips 70° W. The outcrop can be traced 1 mile south from here by the chips of shaly chalk on the surface, but no satisfactory dips could be obtained because of deep weathering, which probably occurred before deposition of the Tertiary, and the extensive cover of gravel and ash washed down from the steep scarp on the edge of the Tertiary beds. On the Niobrara outcrop there are numerous flat fragments of calcite veinlets. Elsewhere these are particularly abundant where the Niobrara is sharply bent. Therefore, a fault is probably covered by the Tertiary a short distance to the west.

The succession from Jelm to Dakota in the scarp at the southwest edge of Red Mountain in sec. 30 is cut by a fault. Its dip is high, and the west block is upthrown. Net slip is 100 to 150 feet. The fault occurs where beds are flexed over the crest of the pre-Cambrian core of Bull Mountain, and it is evidently a tension fracture formed during bending of brittle beds.

Northeast of Red Mountain in secs. 10 and 15 the succession from Casper to Chugwater is sharply bent down to the east in a monocline. Southward the monocline passes into a high-angle fault, and consequently the monocline shown near the east end of section A-A′ (Pl. 1) is shown passing downward into a fault. East of the monocline is a syncline involving beds up to the Dakota; its east flank is cut by a low-angle fault dipping to the west. As the evidence of mode of origin of the monocline, syncline, and two faults is more complete in the area to the south, their genesis is considered later.

BULL MOUNTAIN AREA

General statement.—The name Bull Mountain is applied to three topographically and structurally distinct mountains. The western of these, in which pre-Cambrian rocks are exposed between Red Gulch and Horsepatch Gulch, is anticlinal. The central one extending to Trollope Creek and Shell Creek is synclinal. The eastern mountain is a gentle syncline separated from Green Ridge by the upper part of the valley of Jimmy Creek. The high northeastern part of the eastern mountain is outside the mapped area and adjoins the Sand Creek embayment of the Laramie Basin.

Folds southwest of Bull Mountain.—The rocks in a belt 1 to 2 miles wide southwest of the western part of Bull Mountain are sharply folded. Forester Creek anticline trends almost due north, in contrast with northwesterly trends of folds on both sides. At the south end, at least, the anticline is symmetric and upright. The exposed part plunges south.

Beds in the lower part of the succession are sharply flexed. The Casper in the west flank at the boundary between secs. 7 and 12 dips 50° SW.; in the east flank 800 feet away it is vertical. The sharp fold in beds only 700 feet above the base of the succession strongly suggests that the pre-Cambrian rocks are also folded.

The pre-Cambrian hornblende schists and occasional metadiabase and coarse granite pegmatite forming the core of the anticline appear in an exposure 800 feet long and 300 feet wide around the northeast corner of sec. 12. The schists have excellent cleavage; where frost action has operated on the scarp bordering the river flood plain, the schist is broken into slickensided plates, many of which are less than an inch thick and several inches across. Fractures cutting across cleavage are less numerous than in sharply flexed cross-bedded sandstones and conglomerates of the Casper, Jelm, and Dakota. The metadiabases, which have a poor cleavage, are more extensively fractured and break into irregular angular fragments. The schist contains a few tabular bodies of coarse granite pegmatite up to 5 feet thick. The pegmatites are conformable with cleavage but have not produced extensive *lit-par-lit* injection. Cleavage strikes N. 30°-60°E. and dips 35°-55° NW. The banding of gneisses in the core of Bull Mountain 1000 feet to the east, in contrast, strikes northwest and is nearly vertical. The writer believes that the pre-Cambrian rocks in the core of Forester Creek anticline have been folded along with the overlying sediments.

The Niobrara can be traced almost continuously along the southwest side of the central part of Bull Mountain. In the NE ¼ sec. 18, T. 11 N., R. 76 W., it outlines two sharp folds. Because of lack of space on the map the flank dips of 40°- 60° are omitted; their arrangement indicates that the folds are nearly upright. The syncline in the Niobrara is probably the southeastward-plunging nose of the Laramie River syncline. The exposures of Niobrara about 1 mile southeast in secs. 17 and 20 probably belong to the east flank and crest of the anticline in the NE ¼ sec. 18.

The trend of the belt of folds bordering Bull Mountain on the southwest changes progressively from northwesterly in the southern part to northerly in the northern part; all the folds plunge south to southeast away from the place of tightest folding between the northern part of Forester Creek anticline and Bull Mountain.

Bull Mountain fault.—Southwest of the central part of Bull Mountain beds from Fountain to Benton are involved in a sharp anticline plunging east-southeast. In the NW ¼ sec. 22 a strike fault in the limb between the anticline and the Laramie River syncline omits a small part of the Benton group. Displacement increases to the northwest, and in the SW ¼ sec. 16 Morrison is in contact with middle or upper Benton. Here the fault forks, and the two branches surround a crescentic block of the succession from Chugwater to the Mowry shale of the Benton group. Beds within the block have been turned as much as 45° beyond vertical, indicating that the faults are thrusts dipping northeast. Relations shown in the crescentic block are extensively generalized. Only one tear fault, which crosses from thrust to thrust, is mapped; many others produce offsets too small to map. In addition a number of other thrusts fault out from a few feet to several hundred feet of beds. Unusual fault relations, such as intersecting faults that do not offset each other and tears that offset beds in opposite directions at different places, are present; these are shown diagrammatically and explained elsewhere (Beckwith, 1941b, p. 2185-2187). The branches apparently rejoin to the northwest, and the fault continues beneath alluvial cover.

In the NE ¼ sec. 7 a gulch, not shown on the map, has been cut through the alluvium and Tertiary. Pre-Cambrian rocks are in contact with Mowry shale of the Benton group turned as much as 50° beyond vertical. The fault is not well exposed, and a pit 4 feet deep was dug into it. A northwesterly strike and a dip of 50° NE. were obtained; the dip is undoubtedly a few degrees in error but shows conclusively that Bull Mountain fault is a high-angle thrust dipping northeast.

The Benton outcrop southwest of the main fault is cut by several faults that omit parts of the succession. The Niobrara is excellently exposed in road cuts; it strikes parallel to the mountain front and is slightly overturned. Gray shales are exposed for several hundred feet southwest of the Niobrara. As their age is critical in the interpretation of structure, they were carefully examined. Fossils that would give a positive age determination were not found. The conclusion was reached that the shales are part of the Steele for the following reasons: (1) The gray shales are southwest of the outcrop of the Niobrara, which has a normal width and evidently is not repeated by a fold. (2) The Wall Creek sandstone of the Benton group, which should appear southwest of the Niobrara outcrop if there were a fold in the Niobrara, is absent. (3) The shales weather to a brownish soil characteristic of the Steele; the Carlile shale of the Benton group elsewhere weathers to a gray soil. (4) White calcareous concretions several feet across, characteristic of the Carlile shale, are absent. (5) Within the shales are gray shaly sandstones less than an inch thick, and also flat limonite concretions up to 3 inches across. Both of these are common in the Steele but not in the Benton.

In the NW ¼ sec. 7 the map and section C-C′-C″ (Pl. 1) show, westward from the outcrop of the Steele: fault, Benton, fault, Dakota, fault, Jelm, fault, Chugwater, and an unfaulted succession down to the Fountain. The map is necessarily generalized because of the scale. A traverse westward from the Steele along the meander scarp west of the center of sec. 7 actually shows: fault, middle Benton shales, fault omitting the Mowry shale, upper and middle Dakota, fault, a few feet of Morrison, complete succession of Jelm, and a fault reducing the width of the Chugwater outcrop to about 200 feet. A gulch drains southwest across the middle of the faulted zone from the 65-degree reverse dip symbol on the Niobrara. A traverse westward from the Steele along the north side of the gulch actually shows: fault, Benton with the Mowry absent, fault, upper and middle Dakota, fault, a few feet of Morrison, fault, Jelm without the upper yellow sandstone, Chugwater with width of outcrop reduced to less than 200 feet, and a complete succession down to the Fountain. Beds strike northwest to north and, in most places, are vertical or dip steeply to the east. The Jelm dips 70° W., and cross-bedding shows that the beds are overturned.

The details of succession of outcrops given above are likely to be confusing. Examination, with the aid of the map legend, of the order of outcrops westward from the pre-Cambrian in Bull Mountain to the Fountain in Forester Creek anticline shows: (1) The first three outcrops—Benton, Niobrara, and Steele—are in order from older to younger. This part of the succession consequently belongs to the west flank of the anticline whose pre-Cambrian core forms the western part of Bull Mountain. (2) Westward from the fault bounding the Steele, outcrops are in order from younger to older. This part of the succession consequently belongs to the east flank of Forester Creek anticline. (3) Between the two anticlines there is no syncline; in its place is the fault bounding the Steele.

Relations are believed to be essentially as shown near the center of section C-C′-C″. The eastern branch of Bull Mountain fault produces the common relation of thrusting older rocks onto younger. The western branches produce the unusual relation of thrusting younger rocks onto older. The syncline that is absent at the surface is beneath the pre-Cambrian; inasmuch as Steele shale appears at the surface, beds at least as young as Steele must be present in the core of the syncline. The western main branch, shown underground as a single fault, dips in the same direction as the axial plane of the syncline, but dip of the fault is less than that of the axial plane. The fault fulfills the essential conditions of the type which Billings (1933, p. 147) called an inclined subsequent shear thrust.

The westward dip of the overturned Jelm beds in the east flank of Forester Creek anticline is likely to lead to the conclusion that the fold is overturned and that its axial plane dips west away from Bull Mountain fault. There is no other evidence to support such a conclusion, and the writer believes that the overturning of the Jelm beds was produced as follows: After the beds of the east flank of Forester Creek anticline had been turned nearly to vertical the subsequent shear thrust was initiated at the main fault to the east. The fracture was propagated westward into the anticline and was deflected upward by stratification standing vertically or dipping steeply to the east. Later westward movement of the rocks resting on the curved fault was necessarily rotatory; the block above the thrust was rotated clockwise as seen looking horizontally north at section C-C′-C″, and the Jelm was overturned.

An accurate estimate of the net slip of Bull Mountain fault at the line of section C-C′-C″ cannot be made, for the sediments have been eroded from the block above the thrust, and surface data are inadequate for accurate drawing of the syncline beneath the thrust. The distance scaled on the section from the eastern fault at the surface to the base of the Fountain at the fault is 6500 feet. The distance scaled along the western fault between two points at the base of the Steele is 2300 feet. A conservative figure for total net slip is 8000 feet.

North from Forester Creek anticline Bull Mountain fault is concealed beneath the Tertiary for 2 miles. In the NE ¼ sec. 36 a ridge of pre-Cambrian rocks 800 feet long and 200 feet wide projects through the Tertiary and valley fill. The rocks are almost entirely hornblende schists, greenstones, and metadiabases. Some of the basic rocks are coarse enough to be called gabbros. Several bands of schist up to 30 feet wide have been soaked and injected by granites. All the rocks in the ridge are brecciated, and reliable strikes and dips of schistosity could not be obtained. Sediments from Casper to Chugwater are in contact with the east side of the ridge. At first sight it would seem that the fault is a high-angle normal or reverse fault along which the east wall moved down. Features large enough to show on the map, however, do not confirm this hypothesis. Near the northeast corner of sec. 36 beds from Satanka to Chugwater strike N. 70° E. and dip 15° N. To the west these and the Casper are offset by a small fault, which is apparently a tension fracture formed during bending of brittle beds over the arch at the northwest end of Bull Mountain. Farther west the bands of outcrop on a nearly horizontal surface swing to the south, and dip increases to vertical. If the fault were normal or reverse with the east block downthrown, fault drag should cause the outcrops to swing north and cause the dip to decrease to horizontal and then change to eastward dip in the immediate vicinity of the fault.

Where the Chugwater is shown in contact with the pre-Cambrian a gulch about 5 feet deep has removed the top soil. Here, in order from west to east, appear: (1) Pre-Cambrian rocks. (2) A zone about 4 feet wide of loose brecciated material consisting mainly of gray Morrison shale. The breccia also contains soft yellow sandstone chips possibly derived from the Casper, upper Jelm, or a sandstone in the Morrison. A streak of coal-black shale several inches wide found in the breccia cannot be older than lower Benton, for shales of this color do not occur lower in the succession. (3) Chugwater red shale extends to the outcrop of the Forelle.

Farther southeast, where the map shows pre-Cambrian against Casper, a trench 4 feet deep was started on the pre-Cambrian and extended 15 feet northeast. The trench showed, in contact with the pre-Cambrian, a zone 10 feet wide of plastic black clay. The clay is definitely not a gouge of pre-Cambrian material because it is coal black and contains no breccia fragments. Its contact with brecciated pre-Cambrian greenstones is sharp. The clay is at least as young as Benton, for black shales do not occur lower in the succession. In the northeast end of the trench Chugwater red shale was encountered. Beyond here the surface soil is red for about 30 feet to the outcrop of the Casper. The contact of the pre-Cambrian and black shale was cleaned off over a width of 1 foot and a height of 2 feet. A strike of N. 70° W. and dip of 45° NE. were obtained. These are undoubtedly a few degrees in error because of the small surface exposed; they show, however, that the fault is a high-angle thrust dipping northeast.

The common relation for normal, reverse, and most thrust faults is that breccia fragments and slice blocks within the fault zone at a given place are intermediate in age between the rocks that form the fault walls at that place. However, the breccia fragments and slice blocks in the fault zone in sec. 36 are younger than the fault walls. The trench shows an Upper Cretaceous slice block between fault walls consisting respectively of pre-Cambrian and Pennsylvanian rocks; the beds from which the black shale block was sliced occur at least 1600 feet stratigraphically above the Pennsylvanian. Two thrusts with younger slice blocks have been previously described (Beckwith, 1938, p. 1532, 1541); in both cases the younger slice blocks have clearly been produced by a subsequent shear thrust along which an anticline moved across the adjacent syncline and onto the succeeding anticline. This interpretation is shown near the center of section B-B′; it is thoroughly consistent with relations deduced 2 miles south and shown near the center of section C-C′-C″.

Data from which the central part of section B-B′ was drawn are meager. The slice block of Benton at the surface shows that beds of this age are present in the core of the syncline. To obtain a conservative figure for net slip of the fault, it was assumed that the beds are nearly vertical in the flanks of the syncline. Net slip of the thrust obtained by scaling along the fault from the base of the Casper below the fault to the same horizon above the fault is 6800 feet.

The Dakota in sec. 25, T. 12 N., R. 77 W., changes strike and dip as do older beds to the south, indicating that there is a fault to the west. Bull Mountain fault probably continues beneath the Tertiary for 3 miles and is the one to which the name Ring Mountain fault is applied farther north.

Fault northwest of Bull Mountain.—In the NE ¼ sec. 36, T. 12 N., R. 77 W., the distance from the Casper at Bull Mountain fault to the pre-Cambrian core of Bull Mountain is insufficient for the full thickness of the Fountain. North of the center of sec. 31 space between the Casper and pre-Cambrian is also insufficient. These observations indicate that a fault striking close to east is concealed beneath the alluvium in the lower part of Red Gulch. The fault is probably a tear genetically related to Bull Mountain thrust and consequently an oblique-slip fault with a large component of movement parallel to fault strike. Strike of fault is also close to that of the beds. The south block is upthrown, and stratigraphic separation, or the thickness of beds faulted out, is around 200 to 500 feet. Net slip is, however, probably much greater, for strike slip on a strike fault produces no stratigraphic separation (Beckwith, 1941b, p. 2183).

The fault at the north end of Bull Mountain dies out eastward, for the Casper to the east is not offset. The sharp flexure near the center of sec. 29 in beds from Fountain to Chugwater indicates, however, southwestward movement of rocks south of the flexure. Genetic relations of fault and flexure are considered later in connection with deformation of the western part of Bull Mountain.

Stink Creek fault.—On the east side of the pre-Cambrian core of Bull Mountain in the NE ¼ sec. 8 the Satanka and Casper are in fault contact with granite; the east side of the fault is downthrown, and the beds are sharply dragged up toward the fault. To the south there is no apparent stratigraphic evidence of faulting, as different horizons in the Fountain cannot be distinguished. The straight mountain front and upward drag of the beds immediately to the east indicate, however, that the fault continues south and probably terminates against Bull Mountain fault. On the divide at the head of Horsepatch Gulch, Fountain is in contact with Chugwater. At the head of Stink Creek, so named by local ranchers because of sulfur springs, the succession from Casper to Forelle strikes at about 45° to the Chugwater and younger beds. At the northwest corner of sec. 27 the Forelle is nearly horizontal. It swings down sharply to the east and is cut off by a fault, as the width of outcrop of the vertical Chugwater is about 100 feet less than the thickness of the formation. North of the State line the Chugwater and Forelle are bent down to the east in a monocline. In secs. 15 and 10 the outcrop of the Chugwater is wide enough to represent the whole formation. The fault apparently dies out in the monocline somewhere north of the State line.

Dip of Stink Creek fault could not be obtained, but beds dragged to vertical indicate that it is high. The surface trace of the fault is slightly convex eastward where it crosses the high divide between Horsepatch Gulch and Stink Creek and convex westward in the broad valley about half a mile south of the State line, possibly indicating that the fault dips west. Curvature of the surface trace of the fault is, however, more likely caused by minor changes in strike. In section A-A′ to C-C′-C″ the fault is shown with a high westward dip, so that the fault plane is radial to the arch of the beds over the core of Bull Mountain.

The greatest stratigraphic separation produced by the fault, about 800 feet, is near the middle of Horsepatch Gulch where the Fountain and Casper are omitted. Southward from here the Fountain has been eroded from the upthrown block, and consequently stratigraphic separation cannot be determined. Northward stratigraphic separation decreases to about 200 feet at the head of Stink Creek. The vertical shift (Reid *et al.*, 1913, p. 173), or total vertical movement by combined faulting and bending of rocks, is much larger, possibly as much as 1000 feet as shown in section B-B′.

Deformation around western part of Bull Mountain.—The southern and eastern parts of the pre-Cambrian core of Bull Mountain consist almost entirely of coarse massive granite and gneissoid granite. The western border north from sec. 7 and the north end in the western part of sec. 31 consist of granite gneisses, injected schists, metadiabases, granite pegmatites, and quartz veins. Tabular elements are almost vertical, and northwesterly trends are clear in the aerial photographs. The pre-Cambrian rocks exposed in the core of Forester Creek anticline around the southwest corner of sec. 6 are mainly hornblende schists. They are of a type that is likely to deform by folding with slipping along cleavage; strike of cleavage is northeast, and dip is about 45° NW. The pre-Cambrian rocks projecting through the Tertiary northwest of Bull Mountain in sec. 36 are mainly hornblende schists and greenstones. A discussion of the effect of lithology and structure of the pre-Cambrian rocks on Laramide structure is hampered by lack of data on the crystallines where most needed. The relations described, however, indicate that the granites of the southern and eastern part of the core of Bull Mountain are part of a cupola, and that the gneiss zone of the northwestern border and north end represents a transition to schists farther west and northwest in the zone of sharp Laramide folds.

The western part of Bull Mountain is a northeast-trending anticline. The straight surface trace of the base of the Fountain, the trapezoidal outcrop of the Fountain, and triangular outcrop of the Forelle show that the anticline is flat on top. The flat-topped part is bounded on the southwest by a northeast-dipping thrust fault with a displacement of nearly 2 miles and on the northwest and southeast by high-angle faults. The block enclosed by the faults has moved up along all three. The thrust fault and the sharp flexure outlined by beds from Fountain to Chugwater near the center of sec. 29 indicate that the block has moved southwest with respect to rocks on three sides.

The following hypothesis is advanced to account for described relations. Laramide pressure was close to east-west, as indicated by the general northerly trends of the Front Range, Laramie Range, and the part of the Medicine Bow Range to the north (Fig. 1), and by the northerly trends of folds in the upper Laramie River Valley except in and around the core of Bull Mountain. In the initial stages of deformation rocks now forming the western part of Bull Mountain were gently arched into a northeast-trending flat-topped anticline coinciding approximately with a granite cupola in the pre-Cambrian basement. Stink Creek fault and the one northwest of Bull Mountain were formed as tension fractures generally radial to the arch as shown in section B-B′. Under more intense pressure, folding started in the schist belt to the west, and a southwest-facing step fold developed in the margin between the cupola and schist belt. A thrust fault striking parallel to the face of the step and dipping to the northeast and east was initiated in the massive pre-Cambrian rocks. Dip of the thrust in the massive rocks was less than 45°, in accordance with Hartmann's law (Griggs, 1935, p. 124-126). The thrust was propagated westward into steeply dipping or vertical beds and was deflected upward by stratification.

In depth the thrust cut the older tension fractures. The block between Bull Mountain thrust and Stink Creek fault (sec. C-C′-C″) is a wedge bounded below by a fracture that is concave upward. Westward movement of both blocks above the thrust would necessarily cause the wedge to move up along both faults.

The fault at the north end of Bull Mountain was also cut by the thrust, and the part that now reaches the surface terminates downward on the thrust as shown in section B-B′. Movement on the small fault paralleled that on the thrust. Net slip of the thrust south of the small fault is slightly greater than to the north, producing upthrow of the block south of the small fault.

Bull Mountain syncline.—The central part of Bull Mountain is a broad gentle syncline. The Dakota conglomerates and sandstones form high rims on both sides. The western rim adjacent to Horsepatch Gulch and Stink Creek is 400 to 800 feet high; the eastern is about 300 feet above the valleys of Trollope Creek and Shell Creek. At the south end of the mountain a more gradual slope descends toward the floor of the Laramie River Valley. The cap of the broad southern part is mainly Mowry shale. Beds higher in the Benton may be present beneath the thick pine forest and extensive alluvial cover of Dakota sandstone blocks. From the broad southern part of the mountain a pronounced syncline trends north for about 5 miles and dies out in sec. 15. The fold is nearly horizontal. In most places lower Benton and Dakota beds form the core; Stink Creek has cut through them and as far down as the upper part of the Jelm.

Frenchwoman Creek fault and associated folds.—In the southern edge of the central part of Bull Mountain the Dakota and Morrison are involved in two minor folds trending northwest across the valley of Frenchwoman Creek. They are apparently minor plications genetically related to the larger anticline to the south.

The north flank of this anticline is cut by a fault striking east-southeast across secs. 15, 22, and 23. Exact location of the fault as shown on the map may be, in places, several hundred feet in error because of the extensive debris derived from the Dakota outcrop. North of the fault the position of the base of the Dakota is clearly at the base of the scarp formed by the lower Dakota conglomerate. South of the scarp Morrison variegated shales are exposed in several gulches, and farther south Mowry shale is exposed near Trollope Creek, and Frontier shale near Frenchwoman Creek. Beds on both sides of the fault dip north. The fault dies out near the west boundary of sec. 15, as the base of the Dakota farther west is not offset. It passes beneath alluvium to the southeast and is apparently joined by a small north-striking fault that offsets the lower beds of the Dakota north of the center of sec. 23.

The north side of Frenchwoman Creek fault is upthrown, suggesting a northeast-dipping thrust. The fault, however, is in the northeast flank of an anticline, and vertical to overturned beds were not observed in the vicinity. Evidently Frenchwoman Creek fault was formed in an early stage of deformation by tension in the outer part of a sharp flexure, as shown near the center of section D-D′ in which the fault is radial to the major arch. It is believed that compression in a later stage of deformation caused the hanging wall to move up and also produced the anticline southwest of the fault and the plications to the northeast.

Shell Creek fault.—The east flank of the northern part of Bull Mountain syncline is cut by a fault striking generally parallel to strike of beds. In sec. 35 the fault is locally in the west flank of a gentle anticline. The west block is everywhere upthrown. The largest stratigraphic separation, about 200 feet, is shown several hundred feet north of sec. 35, where the base of the Morrison in the west block is in contact with upper Morrison. The fault outcrop swings markedly to the west in crossing valleys, indicating a low westward dip. Where the Jelm and the Morrison are in fault contact in the NW ¼ sec. 35, a trench was dug down to the fault, and an area 1 foot wide and 2 feet long in the direction of dip was exposed; the fault here dips 25° W., only a few degrees more than the beds. In the NE ¼ sec. 22 a steep-walled gulch exposes the fault for 20 feet normal to strike. Fault dip is 20° W.; the beds dip 10°-15° W. The fault is nearly parallel to stratification, as shown near the east ends of sections A-A′ to C-C′-C″, and closely approaches the conditions necessary for the application of the term bedding fault (Reid *et al.*, 1913, p. 170), a type that produces no offset, repetition, or omission of beds. Net slip of the fault is consequently much larger than stratigraphic separation.

Shell Creek fault is genetically related to competence of beds. The Jelm and Dakota consist mainly of sandstones and conglomerates in which bedding-plane slippage is hindered by cross-bedding. The underlying Chugwater and overlying Benton are thin-bedded shales; gypsum beds in the lower Chugwater and bentonite seams in the lower Benton indicate that bedding-plane slippage is particularly easy in these parts of the succession. Folding of an upright horizontal syncline involves bedding-plane slippage varying from zero at the axial plane, where beds are horizontal, to a maximum in the steepest parts of the flanks. A competent bed in the succession glides outward from the axial plane over incompetent beds. In Bull Mountain syncline, westward gliding of the stiffer plate of Jelm and Dakota beds in the west flank was offset by eastward gliding away from Stink Creek fault and the associated monocline. Gliding of the plate outward from the axial plane of the syncline was thus localized in the east flank. The resulting force couple acting nearly parallel to bedding at this place fractured the competent beds, and a fault was propagated both to the east and west into the shales. The map and sections show that Shell Creek fault has the greatest displacement where Bull Mountain syncline is sharpest and dies out where curvature of beds in the syncline becomes less acute.

Anticline and fault in upper Shell Creek Valley.—The valley between the central and eastern parts of Bull Mountain is carved in the crest of a gentle anticline; flank dips are, in most places, 20° or less. The Dakota forms flanking rims, and the Chugwater is exposed in the center. In the NW ¼ sec. 1 Chugwater beds of the east flank steepen to 60° and are in contact with the lower part of the Dakota. The fault apparently dies out northward in the Morrison; southward it forms the boundary between Chugwater and Morrison for about half a mile. Relations farther south are somewhat uncertain, as the valley sides are heavily forested and extensively covered by debris from the steep slopes on the Dakota. The map of the eastern part of sec. 12 was drawn mainly from topography as seen under the stereoscope. The fault apparently dies out somewhere in this vicinity.

The position of the fault in the east flank of an anticline and the marked steepening of the Chugwater east of the crest indicate that the fault is a westward-dipping thrust of small displacement. It is believed that the fault was initiated in an early stage of deformation as a tension fracture radial to the arch in the brittle Jelm and Dakota beds; under later compression the hanging wall moved up, and the fault was propagated downward and deflected toward stratification in the underlying shales, as shown near the east ends of sections C-C′-C″ and D-D′.

UPPER JIMMY CREEK AREA

General statement.—From the northwest shoulder of Green Ridge beds dip steeply to the west toward the Laramie River syncline and at lower angles to the northwest toward the south end of Bull Mountain. The Dakota striking north from Green Ridge to Bull Mountain forms a hogback 500 to 700 feet high. Jimmy Creek has cut a steep notch in the hogback and a broader valley in the less resistant sediments to the east.

Faults in section 1.—The Dakota and Morrison of the southern part of the hogback in sec. 1 are cut by two faults striking northwest. The lower conglomerate of the Dakota is offset, but the faults die out in the Morrison shales and middle Dakota shales. The block south of each of the faults is upthrown. South of the faults the westward dip of the

Dakota increases rapidly toward Green Ridge fault, an eastward-dipping thrust; the Dakota, consequently, has been subjected to torsional warping on a horizontal line extending north. Motion of the south end of the Dakota was counterclockwise as seen looking horizontally north. The evidence indicates that the faults were formed by torsion and correspond to one of the sets of fractures produced in Daubree's experiments (Leith, 1923, p. 40) of twisting a narrow strip of glass on an axis parallel to its length.

West Jimmy Creek fault—The area east of the Dakota hogback is affected by two faults. The western fault appears in the SW ¼ sec. 31, where it repeats the Forelle. Displacement increases to the north, and in the SE ¼ sec. 25 Forelle on the west is in contact with Morrison. Stratigraphic separation here is about 850 feet and presumably remains about the same or increases slightly to the termination of the west fault against East Jimmy Creek fault in the SE ¼ sec. 24. The trace of the western fault is plainly visible in the steep spur in the SE ¼ sec. 25. A dip of 65° W. was obtained by sighting along strike from about half a mile south of the spur. The block east of the fault is downthrown, and beds in it show normal drag upward toward the fault. Beds in the upthrown block, however, show changes in dip that cannot be explained by fault drag. The Dakota north of the center of sec. 36 dips 30° W. As the fault is approached from the west, direction of dip remains westward, but dip increases to 65° in the Forelle and Satanka; here the beds at the surface are parallel with the fault plane, as shown east of the center of section F-F′. If changes in dip of beds in the upthrown block were produced by fault drag, westward dips should here decrease toward the fault, and in the immediate vicinity of the fault beds should dip east. The writer believes that relations were produced as follows.

During an early stage of deformation a monocline was formed east of the Laramie River syncline as shown in sections E-E′ and F-F′. The Jelm and Dakota conglomerates and sandstones near the line of maximum curvature broke under tension, forming a fracture radial to the arch and dipping steeply to the west. Later, compression forced the hanging wall of the fracture to move up, and the fracture was propagated downward and deflected toward stratification in shales. Possibly the unconformity at the base of the Fountain acted as a glide plane Confirmation of the hypothesis that the fault was originally a tension fracture developed on a flexure is given by the fact that the fault dies out rapidly to the south toward secs. 1 and 6, where beds from Fountain to Dakota have a uniform westward dip of about 30°.

East Jimmy Creek fault.—A fault farther up Jimmy Creek strikes northwest from sec. 29 to sec. 24. Its dip could not be obtained, but, judging from the straight fault trace across a valley about 500 feet deep, the dip is high. In sec. 29 and as far as the end of the Forelle in the southwest block near the center of sec. 30, the southwest block is upthrown as shown in section F-F′. Here drag of beds is consistent with stratigraphic evidence of relative movement of blocks; the 50-degree dips near the fault show that beds in the downthrown block are dragged up toward the fault and beds in the upthrown block are dragged down. Stratigraphic separation where Casper on the northeast is in contact with middle Fountain is about 350 feet; it decreases to the northwest to about 200 feet where the top of the Forelle is in contact with the base of the Casper.

In the NW ¼ NW ¼ sec. 30 direction of movement of blocks is in the opposite direction; older rocks in the northeast block are in contact with the fault, and this block is upthrown. Drag of beds is also consistent with this observation, as shown by the change in strike of the base of the Jelm in the northeast block and the 60-degree dip in the lower Jelm; beds in the northeast block are clearly dragged down toward the fault, as shown in section E-E′. Stratigraphic separation here is about 350 feet. Direction of relative movement of blocks is the same in the vicinity of the southeast corner of sec. 24, where lower Dakota conglomerate on the northeast is in contact with middle or upper Dakota beds.

Apparent upthrow of the northeast block at one end of the fault and downthrow of this block at the other end show that at a point north of the center of sec. 30 throw is zero. This suggests that here the fault is a trace-slip fault (Beckwith, 1941b), one on which direction of movement on the fault is parallel with the trace of stratification on the fault. If this were true, a line representing direction of movement on the fault would plunge northwest at only 20°—the dip of the beds—and the major component of net slip would be strike slip. There is, however, no independent evidence from associated faults of such a large strike-slip component, and fault drag shows clearly that offset of beds was produced mainly by dip slip. The evidence indicates that movement on East Jimmy Creek fault was rotatory about a point north of the center of sec. 30; the southwest block was rotated counterclockwise as seen looking horizontally northeast.

The two Jimmy Creek faults can be traced to within several hundred feet of each other in the SE ¼ sec. 24. Northwest from here to Trollope Creek the rocks are covered by landslide debris from the Morrison. It is highly probable, however, that East Jimmy Creek fault continues beneath the cover and emerges as Frenchwoman Creek fault. If the area beneath cover is a dome, the outcrop of the Dakota crossing the west boundary of sec. 24 seems to require that the south end of the dome be cut by a fault with downthrow on the south as on Frenchwoman Creek fault.

The Dakota outcrop in the E ½ sec. 24 north of West Jimmy Creek fault shows no evidence of faulting, and it is concluded that West Jimmy Creek fault terminates against East Jimmy Creek fault. Exposures north and south of the fault junction provide reliable evidence that, west of the junction, Chugwater south of East Jimmy Creek fault is in contact with Morrison or Dakota; the southwest block, consequently, is again upthrown as in sec. 29. The map symbols showing relative up and down movement at various places on Frenchwoman Creek fault and East Jimmy Creek fault suggest that there are two other pivot points, in addition to the one north of the center of sec. 30. These relations are caused partly by the junction of two faults having unequal stratigraphic separations. The block between the two faults in the SE ¼ sec. 24 is downthrown along both. Downthrow of the east fault is less than 350 feet, the thickness of the Dakota. Downthrow of the west fault is more than 800 feet, the sum of thicknesses of the Morrison, Jelm, and half of the Chugwater. West of the fault junction downthrow of the northeast block, therefore, must be at least 450 feet.

If relations beneath the cover in sec. 24 are essentially as indicated on the map, there must be a second pivot point near the west boundary of the section. Rotation on this point was in the same direction as on the pivot north of the center of sec. 30; the southwest block was rotated counterclockwise as seen looking horizontally northeast. The same counterclockwise rotation is shown by the block west of West Jimmy Creek fault in sections E-E′ and F-F′.

Evidence has been presented to show that Frenchwoman Creek fault and East Jimmy Creek fault constitute a single fault. The writer believes that the fracture was formed in an early stage of deformation by tension in the upper part of a flexure. The southeastern part of the northeast block moved down by gravity. The hanging wall of the northwestern part of the fracture moved up under compression in a later stage of deformation.

SOUTHWESTERN AND SOUTHERN AREA

Laramie River syncline.—Southwest of Bull Mountain and west of Green Ridge is a syncline. The mapped location of the surface trace of its axial plane in T. 11 N. is probably in error because of inadequate exposures. The available data indicate that the fold trends southeast and plunges at a low angle in this direction. The exposures of Shannon sandstone in sec. 34 show that the syncline is an open fold. To the south trend changes to south-southeast, plunge in this direction remains low, and the fold becomes sharper. In sec. 11, T. 10 N., R. 76 W., Mesaverde sandstones close to the axis have dips up to 70°. Dips in the Steele shale and Shannon sandstone to the east vary markedly in short distances, indicating that the beds are drag folded as shown in section G-G′. In the next 3 miles to the south the trend changes to south, and plunge decreases to zero. The east flank is locally overturned. Cross-bedding in the Mesaverde in the SE ¼ sec. 14 shows that the beds are inverted and are part of the east flank of the syncline. Dips in the Mesaverde in the NW ¼ sec. 36 show that the syncline here plunges about 50° N. Older beds of the east flank appear south along the valley of Nunn Creek. The northwesterly dip of the Casper in the SW ¼ sec. 6, T. 9 N., R. 75 W., indicates that the base of the Fountain extends under cover north of the pre-Cambrian in the core of Middle Mountain.

Green Ridge fault.—In the NW ¼ sec. 7, T. 10 N., R. 75 W., Chugwater red sandy shales are exposed in a road cut about 500 feet from pre-Cambrian rocks. Although there is no cross-lamination to show relative ages of beds, it is certain that the red shales are overturned beneath an eastward-dipping thrust. Fault displacement is only a few hundred feet, and the fault apparently dies out a short distance to the northeast. Southwest from the Chugwater exposure displacement increases, and younger beds in the hogback are faulted off. At the north boundary of sec. 12 the distance from Dakota to Niobrara is too small to accommodate the full thickness of the Benton. A branch thrust apparently faults out a few hundred feet of beds and dies out to the north in sec. 1, where the Benton succession is complete. The outcrops of the Shannon and Niobrara are only 1700 feet apart in the SE. ¼ sec. 2, and satisfactory dip observations in the intervening beds could not be obtained. Relations can be explained by steepening of beds to vertical in the narrow interval or by a thrust of small displacement.

West of the center of sec. 13 the pre-Cambrian is in contact with upper Chugwater red shaly sandstones dipping 40° E.; cross-lamination shows that the beds are overturned. The interval of only 1500 feet from the Chugwater exposure to the probable location of the Shannon sandstone beneath cover indicates that there is a thrust of large displacement west of the Chugwater. Half a mile south pre-Cambrian is in contact with Chugwater and Dakota dipping 35° E.; cross-bedding in the Dakota conglomerate shows that the beds are overturned. Three thrust branches are evident here, one between pre-Cambrian and Chugwater, another between Chugwater and Dakota, and the third between Dakota and Shannon. A pit was dug on the eastern branch, and an area of the fault surface 2 by 3 feet was exposed. A strike of N. 10° W. and dip of 35° E. were obtained. The dip is undoubtedly a few degrees in error because of the small area of fault surface exposed but definitely shows that Green Ridge fault is an eastward-dipping thrust.

From sec. 24 to sec. 36 the thrust is covered by talus. Exposures of Mesaverde sandstones to the west and the steepness of the slope between the pre-Cambrian and the road indicate that the Mesaverde extends beneath the cover. Displacement of Green Ridge thrust apparently reaches a maximum somewhere in sec. 25. Net slip as scaled from the land surface to the base of the Fountain beneath the thrust in section H-H′ is 6500 feet, a value undoubtedly less than true net slip because the Fountain has been eroded from the block above the thrust.

Southward the pre-Cambrian is in contact with successively older beds in the east flank of the Laramie River syncline. A thrust branch between Dakota and Niobrara east of the center of sec. 36 can be followed 1 mile to the south. Judging by the distance from Niobrara to Chugwater, displacement does not decrease southward. The branch probably continues up Nunn Creek and offsets the base of the Fountain beneath valley fill and Tertiary beds in sec. 12. The low divide between the heads of Nunn Creek and Porter Creek, which drains south into the Laramie River beyond the boundary of the map, indicates that the fault continues several miles south.

Deadman Creek has cut a valley 500 feet deep in the fault in the SW ¼ sec. 31. The fault can be located within 50 feet in an irrigation ditch south of the creek, and from here a shallow gulch extends south up the main valley wall to a saddle in the crest of a spur. Although the valley wall and spur are heavily covered by pine forest, it is fairly certain that the gulch and saddle are eroded on the fault trace. As observed from a point on the fault about half way up the north wall of the valley of Deadman Creek, the fault to the south dips 40° E. From here into T. 9 N. the pre-Cambrian is in contact with successively older beds, and the fault passes into pre-Cambrian rocks in sec. 7. Although the two branches of the thrust cannot be traced farther south, they probably continue as faults of small displacement beyond the map boundary.

North Middle Mountain fault and associated folds.—At the northeast end of North Middle Mountain in sec. 4 pre-Cambrian massive granites and gneissoid granites are in fault contact with beds from Satanka to Chugwater; the block west of the fault is upthrown. About 100 feet of trench was dug across the fault trace in several places in an unsuccessful attempt to obtain a dip. Even at depths of 8 feet the trenches encountered talus breccia of angular pre-Cambrian blocks in a weathered red clay and white sand matrix derived from Satanka shale and Casper sandstone.

Chugwater and Jelm beds northeast of the fault stand close to vertical. The Morrison is slightly overturned. A branch fault appearing from beneath the Tertiary at the north boundary of sec. 4 cuts out at least 1700 feet of beds between the Morrison and Steele. At the east boundary of the section the fault passes between the Dakota and Niobrara. Farther south a narrow band of Mowry shale appears west of the fault, but here 900 feet of the Benton group is omitted. Beds on both sides of the northeastern fault are overturned as much as 45°, and it is concluded that the faults are branches of a thrust dipping southwest as shown in the western parts of sections F-F′ and G-G′.

The fault can be traced definitely for 3 miles to the northwest. The terrace scarp west of McIntyre Creek in sec. 32 exposes upper Benton beds within 300 feet of the pre-Cambrian. The Wall Creek sandstone is overturned and dips 70° SW. Younger beds appear to the northeast, and the Shannon sandstone outlines a tight syncline trending parallel to the thrust. The fold may be the southern part of one of the synclines that appear 3 to 4 miles north and northwest; it is, however, probably a local fold formed by crumpling beneath the thrust. Exposures of Niobrara and Steele in sec. 31 show that the fault to the southwest has a displacement of several thousand feet. The Steele exposed in sec. 25 in an irrigation ditch 15 feet deep is only 1000 feet from pre-Cambrian rocks, indicating that the fault has a displacement of at least half a mile. Although the shales have no lithologic features that give relative ages of beds, they are probably overturned.

Farther northwest Grace Creek anticline is outlined by beds from Fountain to Chugwater. Steeper dips in the northeast flank show that the axial plane dips southwest; the fold plunges southeast. Beds from Fountain to Chugwater dipping 15°-20° S. in the southwest flank are downthrown against the pre-Cambrian. Although fault dip could not be obtained because of thick talus cover, the fault is probably the northwest continuation of North Middle Mountain thrust as shown near the west end of section C-C′-C″. The Chugwater in the northeast flank of Grace Creek anticline is faulted against younger beds to the northeast. North of Grace Creek a gulch 20 feet deep exposes the fault between Chugwater and Morrison. Four square feet of the fault surface was cleaned off, and a strike of N. 40° W. and dip of 55° SW. were obtained. Beds within a few feet of the fault have the same strike and show dips between vertical and 70° SW. The fault is undoubtedly a thrust of only a few hundred feet net slip, as shown in section C-C′-C″, and is probably a branch of North Middle Mountain fault.

In secs. 4 and 10, T. 10 N., R. 76 W., the branches of North Middle Mountain fault pass southward beneath the Tertiary and flood-plain deposits of the Laramie River. At the southeast end of North Middle Mountain in secs. 26 and 35 a mass of pre-Cambrian gneisses, granites, and pegmatites projects through the Tertiary and stands about 500 feet above the river. The east side of the mass is in fault contact with Steele shale. Trenches up to 8 feet deep failed to expose the fault surface. A pit of the same depth started about 100 feet east of the fault trace where there is no talus cover penetrated brownish-gray sandy clay obviously formed by weathering of the Steele shale but did not reach material fresh enough to show stratification. A pit about 1000 feet from the fault in sec. 26 exposed gray sandy shale striking N. 25° E. and dipping 65° E. toward the Laramie River syncline.

At the south end of the pre-Cambrian knob the fault divides. One branch passes between pre-Cambrian and Chugwater, and the other between Chugwater and Steele. Pits in this vicinity also failed to reach the faults. A pit 5 feet deep several hundred feet from the faults encountered a Chugwater gypsum bed 2 inches thick striking N. 25° E. and dipping from vertical to 80° W. The best evidence of the direction of dip of the fault is provided by the swing of its surface trace. From the vicinity of the Chugwater outcrop the fault trace ascends to the northeast, indicating that the fault dips west. As viewed from the south the dip appears to be about 30° W. This figure is probably less than the dip because the northeasterly strikes of beds indicate that the fault also strikes a few degrees east of north. It is concluded that the fault in secs. 26 and 35 is a westward-dipping thrust, as shown in section H-H′, and is part of North Middle Mountain fault. Net slip here is at least 5500 feet, the distance scaled along the fault from the land surface to the base of the Fountain.

Fold and fault relations.—The area in T. 10 N. between Green Ridge and North Middle Mountain is a structural trough bounded by outward-dipping thrust faults. Southward toward secs. 25 and 26 the syncline between the faults becomes sharper and deeper (secs. G-G′ and H-H′). Although exact data on displacements of faults cannot be obtained because the sediments have been eroded from the upthrown blocks, net slips of faults seem to increase southward; it is at least certain that the pre-Cambrian rocks of both upthrown blocks rest on successively younger beds southward to secs. 25 and 26. Here the distance between the faults at the surface is 5000 feet. Minimum values for net slips of faults scaled from section H-H′ are 5500 feet for the west fault and 6500 feet for the east fault.

As shown by the cross section, one or both of the upthrown blocks must have been eroded during the interval of fault movement in order to permit them to attain their present positions. One fault may be older, in which case extensive erosion must have preceded movement on the other. Movement could have occurred at the same time on both faults if erosion of the upthrown blocks kept pace with fault movement. Stratigraphic data provide no information to indicate that the thrusts are of different ages. Both cut Upper Cretaceous beds. North Middle Mountain fault is covered by Tertiary beds of probable Oligocene and Miocene age. No sediments of undoubted Tertiary age cover Green Ridge fault, but the absence of folds in the Tertiary and the fact that Bull Mountain fault, an eastward-dipping thrust, is covered by the Tertiary beds strongly suggest that all thrusts are pre-Oligocene.

OVERTHRUSTING AND UNDERTHRUSTING

Thrust faults dipping both east and west in the upper Laramie River Valley and adjacent area to the north suggest that a distinction should be made between overthrust and underthrust faults.

Reid *et al.* (1913, p. 179) define overthrusts as reverse faults with low dip or large hade and state that in some cases the dip slip amounts to tens of kilometers. They use the term overthrust with the implication of knowledge only of relative movement of fault blocks.

The discussion of absolute direction of movement on low-angle faults is treated by Hobbs (1914) largely in terms of movements of continents and ocean basins during an interval including a large part of geologic time. He uses the terms overthrust, underthrust, overturned, and underturned in connection with the conclusion that the border of the Pacific Ocean moved toward and under the Asiatic continent.

The terms overthrust and underthrust have been used extensively in connection with pressure box experiments. Link (1928) applies the term overthrust to a fault that dips toward the push block and underthrust to one that dips in the opposite direction. These are valid if the box remains stationary and the push block moves. If, however, the push block were stationary and the rigid box and confined sheet of artificial sediments were moved against the push block, the use of overthrust and underthrust should be reversed. An overthrust fault in Link's sense is essentially one that dips toward a more rigid mass represented by the push block. His findings that faults dipping in this direction are formed about twice as frequently as those dipping in the opposite direction are of considerable interest in explaining the mode of origin of mountain ranges bounded on both sides by inward-dipping thrust faults and fault troughs bounded by outward-dipping thrusts.

Lovering (1932, p. 651) designates as an underthrust a low-angle fault along which the foot wall moved forward actively under a hanging wall that remained nearly stationary. An overthrust fault implies an active hanging wall moving forward over a passive relatively stationary foot wall. He states:

> "If a thrust fault breaks from an overturned fold in a zone of tear faulting, the movement of the walls of the tear faults shows the direction of movement of the adjacent thrust block and thus indicates whether underthrusting or overthrusting has taken place. Similarly, a marked swing of formations toward or away from the axis of the overturned fold as a thrust fault is approached suggests underthrusting or overthrusting respectively."

Willis (1935, p. 413-415) differentiates overthrusts and underthrusts on the basis of location of the adjacent belt of sharp folding. If the rocks above the fault are folded it is an overthrust; if sharp folding occurs below, it is an underthrust.

Attempts to distinguish overthrusts and underthrusts in the upper Laramie River Valley lead to anomalous results. In approaching Bull Mountain thrust fault from the southeast, the outcrop of the Niobrara and the axial planes of the Laramie River syncline and Forester Creek anticline swing toward Bull Mountain fault. On the basis of swing of formations and folding beneath the thrust, the fault is an underthrust. The fault block forming the western part of Bull Mountain is upthrown along faults at both ends. Inasmuch as it is bounded below by a northeastward-dipping thrust, the central block must have moved farther southwest than the bordering blocks. Relations are essentially those illustrated by Lovering (1932, p. 652, Fig. 1) for an overthrust fault.

Where pre-Tertiary beds are exposed near North Middle Mountain fault they strike generally parallel to the fault. The outcrop of the fault and the surface trace of the axial plane of the Laramie River syncline are both convex eastward. Swing of formations suggests an overthrust. The Laramie River syncline beneath the southern part of the thrust and the sharp syncline on McIntyre Creek in sec. 32 suggest an underthrust.

The surface trace of the Laramie River syncline is convex toward Green Ridge fault, and beds in the east flank strike toward the fault, indicating that it is an underthrust. It is difficult to understand, however, why observed relations could not be produced by westward movement of an active east block over a stationary west block.

In view of the inconsistent results the writer has avoided the terms overthrust and underthrust in description and discussion of genesis. The term thrust fault implies only knowledge of relative movement of fault walls and a genetic relation with horizontal compression. Overthrust and underthrust are necessary terms in dealing with present earth movements that can be measured with respect to a fairly stable datum, such as sea level or points outside the zone of movement, and in discussions such as that of Hobbs, which deals with fundamental causes of earth movements. When applied in the study of an area of a few thousand square miles of crust that is now stable, the terms imply a greater knowledge of absolute direction of movement than is commonly justified.

FAULT TROUGH RELATIONS

The upper Laramie River Valley is a structural trough about 20 miles long and 1 to 5 miles wide. It is bounded along most of its length by faults. Those on the east side are clearly outward-dipping thrusts. Although the dip of the fault along the southwest side could not be measured, evidence has been presented to show that it also dips outward. Where the trough is not bounded by faults, beds dip toward it at angles of 30° or more. The base of the sedimentary succession is as much as 8000 feet below erosion surfaces beveling pre-Cambrian rocks at the trough margins. The cross sections show that the trough is deepest where the bordering faults have the greatest displacements.

The Laramie River trough clearly was formed by compression. Although it is small in comparison with those which Taber (1927, p. 591-606) groups under the name profound fault troughs, it may help toward an understanding of the genesis of some of the large troughs whose border faults are covered. Lakes commonly occur close to the edges of large troughs, and the deepest parts of submarine troughs are commonly in the same position. These facts have been advanced as evidence that the trough was not formed by compression. Sections A-A′ to D-D′ show that the deepest parts of the trough are adjacent to thrust faults of large displacement. Marginal deepening could be produced either by downwarping of the border under the weight of an advancing thrust block or by unparching of the central part of the trough under compression.

CONCLUSIONS

The general northerly trends of the Front, Laramie, and Medicine Bow ranges (Fig. 1) and of the larger anticlinal mountains along the west margin of the Laramie Basin indicate that deformation was caused by east-west compression with predominant upward relief. A laterally uniform sheet subjected to pressure in this direction develops northward-trending folds and northward-striking thrust faults. The structural pattern of the upper Laramie River Valley differs markedly from this ideal pattern because of heterogeneity of the crystalline rocks. The Laramie River trough varies in trend between north and northwest in accordance with the strikes of the bordering thrust faults. Folds within the trough and to the northeast trend from northwest to northeast.

The predominantly shaly Paleozoic and Mesozoic sedimentary succession shows little facies variation and can be considered as laterally uniform. The beds as a whole were incompetent and deformed mainly in response to movements of the crystalline rocks. Thin conglomerate and sandstone beds, however, were strong enough to support at least part of the load on an arch and caused local minor folding. They also fractured by tension where sharply flexed and initiated faults along which movement occurred during a later stage of compression. The pre-Cambrian igneous and metamorphic rocks vary markedly in character in short distances, and the writer believes that this is a major cause for the divergence of the observed structural pattern from one produced by compression of a laterally homogeneous sheet. In one area of complex structure where the crystalline and sedimentary rocks are both exposed, observations indicate that schists in the pre-Cambrian have been folded and an adjacent granite mass has deformed by thrust and tear faulting. Stratigraphic data in adjacent areas show that erosion had locally cut away about 2 miles of sedimentary cover before the end of the interval of deformation. Depth of burial influences the manner in which a given rock type deforms, and consequently local stripping of the sedimentary cover has undoubtedly had some influence on the structural pattern.

Application to the thrusts in the upper Laramie River Valley of criteria that attempt to distinguish absolute direction of movement on a thrust fault lead to inconsistent results. The criteria deal essentially with differential crustal shortening by folding and by movement on thrust and tear faults. Vertical and horizontal variation in sharpness of folds, variation in trend of folds and strike of beds, and variation in net slip of tears and thrusts are all necessary consequences of application of compression to a laterally and vertically nonuniform crust. The fact that such variations are observed in certain associations does not distinguish an active from a passive fault block; the observed results could be produced by equal differential movements irrespective of absolute direction of movement of hanging wall or foot wall.

Pressure box experiments do not provide evidence of absolute direction of movement of thrust walls. They show that thrusts most commonly dip toward the push block, which is a rigid mass separated below and at the ends from the floor and lateral walls of the box. The outward-dipping thrusts bounding the Laramie River fault trough suggest an analogy with a pressure box provided with push blocks at both ends. The push blocks are represented by Bull Mountain and Green Ridge at one end and the Medicine Bow Range and North Middle Mountain at the other.

These masses functioned as push blocks either because they were more rigid than the rocks between or because they were upwarped in an early stage of deformation before the thrust fractures were formed.

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RING STRUCTURES OF THE PLINY REGION, NEW HAMPSHIRE

BY RANDOLPH W. CHAPMAN

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ABSTRACT

The Pliny region, in the northern portion of the Mt Washington quadrangle, New Hampshire, contains some of the finest examples of ring dikes described in North America. These arcuate bodies intrude older gneiss, quartzite, and quartz diorite, and are, in turn, cut by later granitic stocks The dikes and stocks, which are grouped about two distinct centers, are composed of differentiates of the White Mountain magma series, which in this area range from quartz monzodiorite to granite. Two of the ring dikes are strikingly expressed in the topography as arcuate mountain ridges

All the ring dikes and some of the stocks are thought to have originated by cauldron subsidence or ring-fracture stoping However, this mechanism operated differently for different dikes In a composite ring dike in the southwestern part of the area, an arcuate zone of intense fracturing developed, resulting in the subsidence of a large cylindrical or domical block of country rock Numerous small dikes of hastingsite-quartz syenite penetrated the northern part of this zone to form a more or less solid dike of hastingsite-quartz syenite Later, pink biotite granite intruded the southeastern part of the fractured zone, forming a network of irregular dikes.

The Crescent Range ring dike of granite porphyry, however, is a symmetrical crescent with regular, broadly sweeping boundaries. This suggests that it was formed by intrusion en masse along a clean, sharp ring fracture The dense groundmass of the rock implies rapid cooling

INTRODUCTION

LOCATION OF AREA

The "Pliny region" covers an area of approximately 100 square miles (Fig. 1), chiefly in the northern portion of the Mt. Washington quadrangle in north-central New Hampshire, between Lats. 44°30′30″ and 44°23′40″ and Long. 71°15′ and 71°30′. The Percy quadrangle (R. W. Chapman, 1935), to which numerous references are made in the present paper, lies directly north. It will be noted, by referring to the latitude lines on the geological map (Pl. 1), that the extreme southern portion of the Percy quadrangle is included in the area which is here known as the "Pliny region."

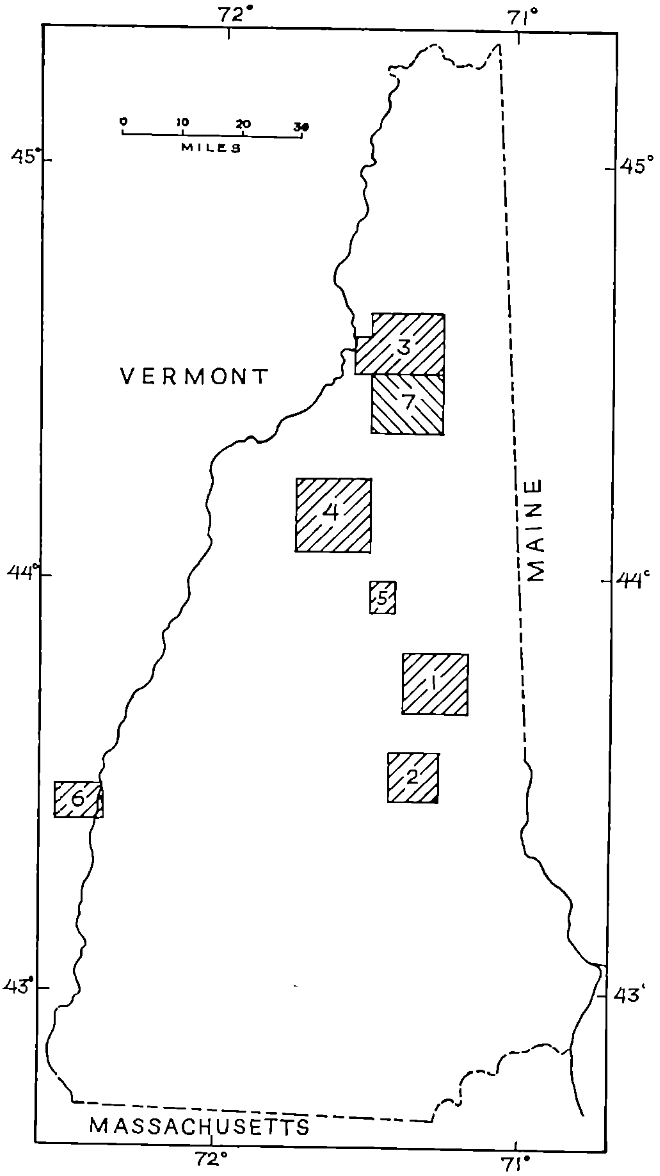


FIGURE 1—*Index map of Pliny region and other ring dike areas*

1 = Ossipee Mountains, 2 = Belknap Mountains,
3 = Percy region, 4 = Franconia region, 5 = Mt. Tripyramid,
6 = Ascutney Mountain, 7 = Pliny region.

GENERAL GEOLOGICAL RELATIONS

The Pliny region contains some of the finest examples of ring structures described in North America. The "complex" consists of gneissic plutonics intruded by arcuate ring dikes and associated granitic stocks. The ring dikes and stocks are grouped about two distinct centers and are composed of differentiates of the White Mountain magma series. On the whole, these intrusives are more resistant than the surrounding gneisses so that they form high mountains. Two ring dikes are strikingly expressed as arcuate mountain ridges known as the Pliny Range and the Crescent Range.

TOPOGRAPHIC FEATURES

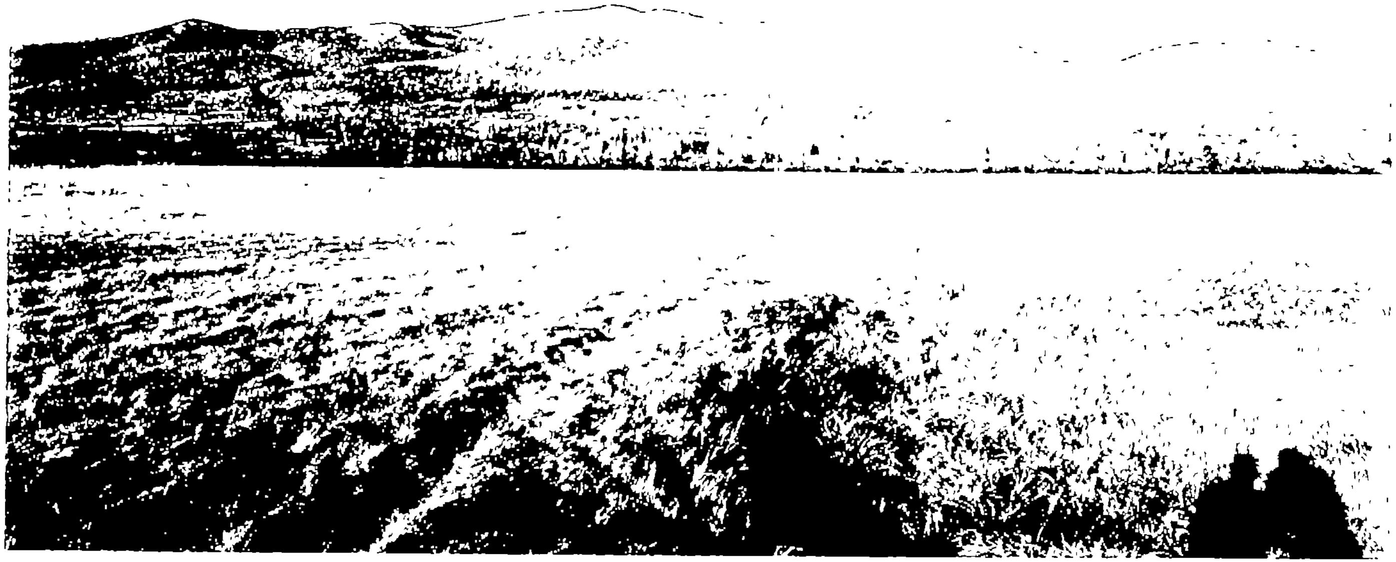
The region considered in this study contains two mountain ranges and a portion of a third. The dominant Pliny Range is distinctly arcuate, has a radius of curvature of 2 miles, and a length of approximately 7 miles. The three highest peaks are Mt. Waumbek (elevation 4005 feet), Mt. Starr King (3913 feet), and Round Mountain (3890 feet). The range takes its name from Pliny Mountain which rises to a height of 3605 feet on the southeast side of the mass (Pl. 2).

Farther to the east is the Crescent Range. Although nearly 1000 feet lower than the Pliny Range, it is no less striking topographically. It is broadly arcuate with a radius of curvature of about 4 miles and a length of over 12 miles. Black Crescent Mountain (3265 feet) and Mt. Crescent (3230 feet) are the two highest peaks (Pl. 1).

North of the Pliny Range a portion of the Pilot Range extends southward into the area. This is represented by the southern slope of Mt. Cabot (4160 feet) and by Terrace Mountain (3640 feet). The topographic break between the Pilot and Pliny ranges is not strong.

The topography and drainage of the Pliny region are determined primarily by differences in rock hardness. The ranges, composed of hard resistant intrusives, rise above the surrounding lowlands which are underlain by older and softer intrusive and metamorphic rocks. The three main streams are the Israel River in the southern half of the area, the Upper Ammonoosuc River in the northeastern portion, and westerly flowing Garland Brook which drains that portion northwest of the Pliny Range. A myriad of tributaries, flowing from the mountains, form an intricate drainage pattern in perfect adjustment to the geologic structure (Pl. 1).

The Pliny region is mainly a wilderness. Except in the extreme western portion, the mountains and lowlands alike are clothed with a dense cover of birch, maple, aspen, beech, spruce, and fir. No roads cross the center of the area so that the interior is accessible only by foot trails.



PLINY RANGE, NEW HAMPSHIRE, AS VIEWED FROM THE SOUTHWEST

Peaks in center are Mt Starr King (left) and Mt. Waumbek. Peak on extreme right is Pliny Mountain. Village of Jefferson at foot of range on left.

Since most of the area lies within the White Mountain National Forest, however, these trails are kept in first-class condition at all times.

GLACIAL DRIFT

The Pliny region is heavily glaciated. Locally the drift completely conceals the underlying rock, and these places have been mapped as glacial drift (Pl. 1). For example, the topographic basin north of Mt. Starr King and Mt. Waumbek is filled with glacial drift which in many places is probably 100 feet or more thick. The flat area south of the village of Jefferson is likewise deeply covered. Somewhat thinner drift occurs northwest and southeast of Mt. Crescent, in the extreme northeast portion of the area, and along the broad valley of the Upper Ammonoosuc River.

METHOD OF STUDY

A number of reconnaissance trips were made by Marland Billings and the writer during the summer of 1936. Detailed mapping, however, occupied approximately 11 weeks during the summers of 1939 and 1940. Enlarged photostatic copies of the Mt. Washington quadrangle of the United States topographic atlas were used as base maps. Elevations were determined by means of an aneroid barometer.

At the time the field work was done, Billings was studying the structure and metamorphism of the Presidential Range. As both areas lie on the same topographic sheet, it was thought advisable to map the entire Mt. Washington quadrangle. Accordingly, Billings undertook the mapping of the southern half of the topographic sheet, and the writer, assisted by Carleton A. Chapman, completed the northern half.

Billings (1941) has published on the structure and metamorphism in the Mt. Washington area, and the present paper deals with the ring structures of the northern part of the Mt. Washington quadrangle. A third paper is in preparation on the petrology and structure of the Oliverian magma series in the Mt. Washington quadrangle. Later, it is hoped that a colored geological map and a quadrangle report may be published.

ACKNOWLEDGMENTS

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NATURE OF THE COUNTRY ROCK

Since the exact nature of the country rock is not pertinent to the present problem, it will be mentioned only briefly here.

The oldest formation, the Albee, occurs as a narrow, east-west band, 1 mile long, 4 miles northwest of Mt. Starr King. It is a fine-grained, gray, impure quartzite and has been described in detail by Billings (1937, p. 472-475). Its age is pre-Silurian and probably late Ordovician. Large inclusions of it are abundant in the hastingsite-riebeckite granite on Terrace Mountain (Pl. 1).

Immediately south of the Albee formation, northwest of Mt. Starr King, are the Ammonoosuc volcanics. This volcanic formation is also pre-Silurian (Billings, 1937, p. 457-480) and stratigraphically overlies the Albee formation in the Littleton quadrangle. Its age is also probably late Ordovician. The volcanics are composed largely of fine-grained biotite gneiss with minor amounts of amphibolite and mica schist.

In the extreme northwest corner of the area the country rock belongs to the Lost Nation group (Chapman, 1935, p. 405) which here consists of medium-grained, dark intrusives, including mainly quartz diorite and diorite with some gabbro and syenite. All these rocks show considerable hydrothermal alteration and metamorphism. The Lost Nation quartz diorite is assigned to the Highlandcroft magma series (Billings, 1937, p. 499-500) near Littleton, New Hampshire, which is believed to be late Ordovician.

Most of the country rock consists of the Oliverian magma series (Billings, 1937, p. 501-502), a group of medium- to coarse-grained, light-colored intrusives probably of late Devonian age. In the Pliny region this series consists of: (1) biotite gneiss, (2) porphyritic biotite gneiss, (3) coarse granite, (4) coarse syenite, (5) hornblende-quartz monzonite, and (6) fine, gray quartz monzonite. Some of these types are distinctly foliated whereas others are massive, but all are of igneous origin. The general distribution may be seen in Plate 1. No attempt has been made here to show the distribution of the varieties. In areas southwest of the Pliny region the Oliverian magma series forms a long line of domelike bodies trending north-northeast. Structural studies indicate that the Oliverian rocks in the Pliny region are also part of a large dome into which ring dikes and stocks have been intruded.

The boundary between the Lost Nation group and the narrow band of Albee formation on the north, and the Ammonoosuc volcanics on the south is probably the northward extension of the Ammonoosuc thrust fault (Billings, 1937). Throughout most of western New Hampshire the Ammonoosuc thrust fault trends generally northeast, but in the Pliny region it apparently strikes west-northwest. Further work west and southwest of the Pliny region is necessary to substantiate this view. The Ammonoosuc thrust fault is older than the White Mountain magma series.

PETROGRAPHY OF THE WHITE MOUNTAIN MAGMA SERIES

GENERAL STATEMENT

The rocks forming the ring dikes and stocks belong to the White Mountain magma series which, according to Williams and Billings (1938, p. 1025), is probably Mississippian or early Pennsylvanian. The individual rock types of this series have been previously described in many reports (Daly, 1903; Eggleston, 1918; Billings, 1928; Kingsley, 1931; Jenks, 1934; R. W. Chapman, 1935; 1937; Chapman and Williams, 1935; Modell, 1936; Quinn, 1937; Williams and Billings, 1938; Smith *et al.*, 1939; Chapman and Chapman, 1940). In addition, a comprehensive study of the whole series has been made by Chapman and Williams (1935). Accordingly, only a brief petrographic description of each type is given here. Modes of all rock types compose Table 1.

Special optical study was made of the minerals of the plutonic rocks, and the results are described fully. No new chemical analyses of minerals are available, but on the basis of optical properties the chemical compositions can be compared with chemical analyses of minerals of the White Mountain magma series (Table 5). It is believed that in the future such optical and chemical properties will be of great value in determining the course of differentiation of the White Mountain magma series.

The intrusive rocks of the White Mountain magma series found in the Pliny region, exclusive of the associated complementary dikes, are as follows, oldest at the bottom:

Conway granite
Pink biotite granite[1]
Granite porphyry[1]
Hastingsite-biotite granite[1]
Hastingsite-riebeckite granite
Hastingsite-quartz syenite
Quartz monzodiorite and quartz monzonite

The age sequence is based upon: (1) relations as determined at contacts between these rock types in the Pliny region, and (2) correlation of these rocks with their equivalents in other regions where the relative ages are known. Age relations are discussed in detail in a later section.

QUARTZ MONZODIORITE AND QUARTZ MONZONITE

The quartz monzodiorite and quartz monzonite cannot be differentiated megascopically. Inasmuch as the two rock types apparently grade into one another, both are shown on the geological map (Pl. 1) by the same

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[1] The exact position in the sequence of these rocks is not certain.

TABLE 1.—*Modes, in volume per cent, of intrusive rocks of Pliny region, New Hampshire*

ROCK TYPE	Specimen Number	Orthoclase	Microperthite	Plagioclase	Quartz	Diopside	Hedenbergite	Hornblende	Hastingsite	Riebeckite	Biotite	Magnetite	Apatite	Zircon	Sphene	Allanite	Fluorite	Astrophyllite	Chlorite (from biotite)	Epidote (from plagioclase)	Hematite
Quartz Monzodiorite	15	5		44(An_{35})	10	4		26			9	2	tr	tr						tr	
	39	12		51(An_{45})	5	1		14			13	2	2		tr						
	78	6		52(An_{35})	3	23		2			5	2	tr		tr				7	tr	
	87	5		69(An_{40})	6	3		10			4	2	1		tr				tr		
	88	23		54(An_{40})	10	2		8			tr	2	tr	tr					1		
	113	15		42(An_{35})	8	3		15			8	3	1		1				4	tr	
	120	16		46(An_{35})	10			15			14	tr	tr	tr	tr				tr		
Quartz Monzonite	10	42		23(An_{40})	17			13			1	2	tr		tr				2		
	38	25		38(An_{30})	10			8			16	2	1	tr							
	85	28		36(An_{40})	9	tr		15			12	tr	tr		tr				tr		
	114	20		41(An_{30})	16			13			10	tr	tr		tr	tr			tr		
Hastingsite-Quartz Syenite	11		66	18(An_{15})	11				5		tr	tr		tr		tr			tr		
	34		80	7(An_{10})	10				3		tr	tr	tr	tr	tr				tr		tr
	35		68	22(An_{15})	8				1		tr	1	tr	tr	tr						
	65		74	3(An_{10})	19		3					1		tr							
	77		80	1(An_{10})	8		7		2			2		tr		tr			tr		
	122		62	19(An_{10})	12				6			tr		tr	tr	1					
	126		72	7(An_{15})	14				5		2	tr	tr	tr	tr	tr					
Hastingsite-Riebeckite Granite	18		61	18(An_{10})	14		tr		6	tr	tr	1	tr		tr	tr	tr		tr		tr
	81		55	3(An_{10})	37				3		2	tr	tr	tr	tr	tr					
	96		50	12(An_{10})	32					6	tr	tr	tr	tr	tr	tr	tr	tr			tr
Hastingsite-Biotite Granite	62		52	13(An_{10})	31				1		3	tr	tr	tr					tr		
	63		68	12(An_{10})	15				1		3	tr		tr	tr				1		tr
	64		51	17(An_{10})	20				11			1		tr		tr			tr		tr
Granite Porphyry	17	40	17	3(An_{10})	39						tr	1	tr		tr	tr			tr		
	31	9	52	3(An_{10})	29				1		4	1	tr	tr	tr	1	tr				
	43	15	49	1(An_{10})	28				tr		6	tr	tr	tr	tr				1		tr
	45	33	31	4(An_{10})	24				5		2	1	tr	tr	tr	tr					
	56	39	26	1(An_{10})	26						6	1	tr	tr					1		
	60	33	27	1(An_{10})	28		1		5		3	2	tr	tr		tr					
Pink Biotite Granite	116		61	14(An_{15})	17						5	2	tr	tr	1	tr			tr		
	127		50	12(An_{15})	37							tr	tr	tr	tr	tr			1		
Conway Granite	36		47	26(An_{15})	24						2	1		tr	tr	tr					
	98		49	23(An_{15})	27						1	tr	tr	tr					tr		

pattern. They form two separate bodies. The larger body lies on the north side of the Pliny Range, but unfortunately, heavy glacial cover conceals its areal distribution and original shape. The smaller body is a ring dike, 1 square mile in area, which lies due north of Jefferson. Furthermore, the quartz monzodiorite occurs both as inclusions and dikes in the shatter zone portion of the composite ring dike northeast of Jefferson. A large inclusion of quartz monzodiorite, too small to show on Plate 1, was found in the hastingsite-riebeckite granite at an elevation of 3200 feet on the southeast slope of the eastern knoll on Terrace Mountain.

The quartz monzodiorite and quartz monzonite are similar texturally and contain the same minerals but in different proportions. The two types are named according to the rock classification that has generally been used for the White Mountain magma series (Chapman and Williams, 1935). Field relations and laboratory studies show that both of these types are closely related genetically and they appear to grade into one another.

The quartz monzodiorite and quartz monzonite are dark-gray and even-grained with varying proportions of light and dark minerals. The grain size ranges from 0.5 to 3 millimeters in different specimens, although the feldspar phenocrysts in one porphyritic specimen are 1 centimeter in diameter. On the whole the dark constituents are somewhat smaller than the light. In most specimens the normal texture is granular, but in a few a diabasic texture is apparent. Some varieties of the rock are pinkish due to the presence of abundant orthoclase.

Microscopically, all these rocks are hypidiomorphic granular and generally even-grained. Some thin sections show a decided diabasic texture. The essential minerals are orthoclase, plagioclase, quartz, pyroxene, hornblende, and biotite. Accessories include apatite, magnetite, zircon, sphene, and allanite. Secondary minerals such as chlorite, epidote, sericite, and kaolin are abundant. Modal analyses of several specimens are shown in Table 1.

Orthoclase makes up from 5 to 23 per cent of the quartz monzodiorite and from 20 to 42 per cent of the quartz monzonite by volume. It occurs both as individual crystals and as borders on plagioclase laths. All grains are altered either to kaolin or to sericite. Lath-shaped plagioclase crystals compose from 42 to 69 per cent by volume of the quartz monzodiorite and from 23 to 31 per cent of the quartz monzonite. Most of it is sodic andesine (An_{30-40}), but in two specimens of quartz monzodiorite it is labradorite (An_{55}). Polysynthetic twinning is common and most crystals show distinct zoning with cores of andesine and borders of oligoclase. Many cores have altered to epidote and sericite. Quartz, in varying proportions (Table 1), is molded into the interstices between the feldspars and ferromagnesian minerals. Pyroxene, hornblende, and biotite are the chief dark minerals.

The pyroxene occurs as irregular cores, about 0.5 millimeter in diameter, rimmed by pale green hornblende, indicating that the hornblende has been derived from the pyroxene by a late magmatic reaction. The pyroxene is nearly colorless to pale pinkish-yellow in thin section, and contains abundant poikilitic inclusions of quartz and feldspar. Some of the pyroxene from specimen 78 of the quartz monzodiorite was isolated for optical study and gave the properties shown in Table 2.

TABLE 2.—*Optical properties of pyroxenes from rocks of Pliny region*

		1	2	3
Indices	α β γ	1.691 1.698 1.718	1.730 1.737 1.757	1.730 1.736 1.755
Sign		Positive	Positive	Positive
2 V		60°	60°	60°
Dispersion		r > v, weak	r > v, medium	r > v, medium
Orientation		$Z \wedge c = 43°$ $Y = b$	$Z \wedge c = 45°$ $Y = b$	$Z \wedge c = 45°$ $Y = b$
Pleochroism		X = pale yellow—pink Y = pale yellow—pink Z = pale blue—green $X = Y < Z$	X = light green Y = light green Z = green $X = Y < Z$	X = dark olive—green Y = dark yellow—green Z = dark green $X = Y < Z$

1. Pyroxene from quartz monzodiorite (specimen 78).*
2. Pyroxene from hastingsite-quartz syenite (specimen 65).
3. Hedenbergite from syenite in the Percy quadrangle, New Hampshire (Chapman and Williams, 1935, p. 512-513).

* Specimen numbers correspond to specimen numbers in Table 1.

According to Winchell's diagram of the diopside-hedenbergite series (1933, p. 226), it is a diopside, and consists of 63 per cent by weight of diopside and 37 per cent by weight of hedenbergite. It contains 12 per cent by weight of MgO and 11 per cent of FeO.

The hornblende in the quartz monzodiorite and quartz monzonite is green-brown and occurs: (1) as distinct crystals, and (2) as reaction rims around pyroxene. The crystals range in diameter from 0.3 to 1 millimeter and average 0.5 millimeter. A few contain poikilitic inclusions of quartz and feldspar. The reaction rims are irregular and vary greatly in width. It is apparent that both the hornblende crystals and rims have resulted from the magmatic reworking of pyroxene.

In different specimens the indices of refraction of the hornblende vary somewhat, although the other optical properties are essentially constant. These relations are shown in Table 3, where the optical properties of two different specimens of hornblende from the quartz monzodiorite (specimens 87 and 120) are compared with those of a hornblende from a syenite of the White Mountain magma series in the adjacent Percy quadrangle.

The refractive indices suggest that the hornblende from specimen 120 has a somewhat higher MgO:FeO ratio than that from the Percy syenite, a chemical analysis (Chapman and Williams, 1935, p. 512) of which is shown in Table 5. The hornblende from specimen 87 would probably be even richer in magnesia. A rechecking of data suggests that the optic angle of the hornblende from the Percy syenite is about 65° rather than 35° as reported by Chapman and Williams.

The optical properties of the hornblende from specimen 87 are the same as those of an amphibole from a diorite of the White Mountain magma series from Mt. Pequawket in the North Conway quadrangle, New Hampshire (Billings, 1928, p. 103). According to Billings the amphibole from the North Conway quadrangle is a normal hornblende relatively rich in magnesia compared to hastingsite.

Brown biotite, abundant in both the quartz monzodiorite and quartz monzonite, occurs as ragged crystals, 0.3 to 1 millimeter across, some of which are intergrown with pyroxene and hornblende. Poikilitic inclusions are not uncommon, and many grains are altered to green chlorite along cleavage cracks. Refractive indices differ slightly in different specimens of biotite. Complete optical data on two specimens of biotite are shown in Table 4. These biotites were chosen from the same two specimens of quartz monzodiorite (specimens 87 and 120) that yielded the two hornblendes already described.

The optical properties of the biotite from specimen 87 are identical with those of a biotite from the Ames monzodiorite in the Belknap Mountains, New Hampshire (Chapman and Williams, 1935, p. 512-513). Thus the chemical composition is probably essentially the same A chemical analysis of the biotite from the Belknap Mountains is shown in Table 5.

The biotite from specimen 120 is optically like the biotite from the Conway granite (see Table 4), which suggests identical chemical composition. However, it seems strange that two rocks, such as the quartz monzodiorite and the Conway biotite granite, whose chemical compositions are so different, should contain biotites which are chemically alike. Possibly the indices are not reliable in determining accurately the chemical composition of the biotite.

HASTINGSITE-QUARTZ SYENITE

The areal extent of the hastingsite-quartz syenite in the Pliny region is between 5 and 6 square miles. This resistant rock makes up the largest ring dike in the region, and is chiefly responsible for the great height of the Pliny Range. Hastingsite-quartz syenite also occurs in three other places in the Pliny region: (1) in the northern part of the composite ring dike north of Jefferson, (2) as a small stock 1 mile southwest of Pliny Mountain, and (3) as a small lenslike mass at the northern end of the ring dike of quartz monzodiorite and quartz monzonite 2½ miles northwest of Mt. Starr King.

In hand specimen the typical hastingsite-quartz syenite is even-textured to subporphyritic, medium-grained, and composed essentially of feldspar and hastingsite. A few specimens, however, are medium-fine grained and distinctly porphyritic. The feldspars are commonly lathlike, and are locally arranged in a trachytic fashion. They are 2 to 5 millimeters long and 1 millimeter wide. In other specimens the feldspars are equidimensional with an average diameter of from 1 to 2 millimeters. Carlsbad twins are common. The dark minerals range from a fraction of a millimeter to 3 millimeters. The fresh rock is bluish green but weathers white, pink, or rusty brown.

Microscopically the hastingsite-quartz syenite is hypidiomorphic-granular and consists essentially of microperthite, plagioclase, quartz, and hastingsite. For the most part it is even-grained, but a few specimens are seriate. Accessory minerals are pyroxene, biotite, apatite, magnetite, zircon, sphene, and allanite. Chlorite (from biotite) and hematite are secondary. Modes of several specimens of hastingsite-quartz syenite are shown in Table 1.

The microperthite consists of potash feldspar penetrated by long, slender, parallel stringers or irregular patches of albite. The ratio of orthoclase to albite by volume in the microperthite is from 3:1 to 2:1. Adjacent perthite crystals are commonly intergrown along sutured or interdented boundaries. The plagioclase is sodic oligoclase (An_{10}-An_{15}) and occurs generally as small anhedra showing albite twins. Quartz occurs chiefly interstitially between the feldspars. Table 1 shows that the quartz content ranges from 8 to 19 per cent by volume. However, most specimens contain between 8 and 15 per cent and are thus classified as hastingsite-quartz syenites.

The hastingsite in the hastingsite-quartz syenite is dark green to olive green. Most grains are subhedral or anhedral, 0.3 to 3 millimeters in cross section, and dotted with poikilitic quartz and feldspar. Many of these grains are interstitial between the feldspars suggesting late crystallization. A small percentage of the hastingsite occurs as irregular masses altering from hedenbergite. The optical properties of the hastingsite from specimen 34 are shown in Table 3, where they may be compared with those of a hastingsite from a quartz syenite in the North Conway quadrangle, New Hampshire.

Billings (1928, p. 110) gives a chemical analysis of the North Conway hastingsite and this is shown in Table 5. Inasmuch as the refractive indices of the hastingsite from the hastingsite-quartz syenite of the Pliny region are slightly higher than those of the North Conway hastingsite, it may be that the hastingsite from the Pliny region is poorer in magnesia and richer in ferrous iron.

Pyroxene is not abundant, but it was observed in a few sections as small, pale-green to bright-green, subhedral or euhedral crystals 0.1 to 0.5 millimeter across. Many crystals show fractures along which the pyroxene has altered to yellow antigorite. In Table 2 the optical properties of this pyroxene, taken fom specimen 65, are compared with those of a hedenbergite from a syenite in the Percy quadrangle, New Hampshire.

These two pyroxenes are essentially the same optically except that the hedenbergite from the Percy quadrangle is somewhat more deeply colored. The chemical compositions of the two are probably about the same, i.e. both minerals are hedenbergite. The hedenbergite from the syenite of the Percy quadrangle has been analyzed chemically (Chapman and Williams, 1935, p. 512) and this analysis is shown in Table 5.

HASTINGSITE-RIEBECKITE GRANITE

Hastingsite-riebeckite granite forms an irregular stock-like mass along the northern boundary of the area. Its areal extent in the Pliny region is approximately 6½ square miles. This rock type is an extension of a hastingsite-riebeckite granite body in the Percy region (Chapman, 1935, p. 426-428). There the rock contains three different amphiboles: hastingsite, riebeckite, and hornblende. In the Pliny region, however, hornblende is absent. Inasmuch as some specimens from the Pliny region contain both hastingsite and riebeckite it was found impracticable to try to separate this amphibole granite into two distinct types.

The hastingsite-riebeckite granite is medium-grained and consists principally of feldspar, quartz, hastingsite, and riebeckite. Its texture, both megascopically and microscopically, is similar to that of the hastingsite-quartz syenite, and the color of the hastingsite granite phase is exactly like that of the hastingsite-quartz syenite. Some specimens of the riebeckite phase are white or buff and dotted with jet-black amphibole.

The hastingsite-riebeckite granite is generally richer in quartz than the hastingsite-quartz syenite. Microperthite is somewhat less abundant, and in it the ratio of potash feldspar to soda feldspar is slightly less than in the microperthite of the hastingsite-quartz syenite. Sodic oligoclase (An_{10}) is present in about the same amount as in the hastingsite-quartz syenite. Some specimens contain only hastingsite, some only riebeckite, and others contain both. Accessories include astrophyllite, allanite, fluorite, zircon, sphene, hedenbergite, biotite, apatite, yellow chlorite, magnetite, and hematite. Modes are shown in Table 1.

Most of the hastingsite occurs as clean, irregular crystals, 1 or 2 millimeters across. Its interstitial occurrence between subhedral feldspars suggests late crystallization. Some hastingsite forms alteration rims on bright green hedenbergite.

The optical properties of the hastingsite from specimen 18, shown in Table 3, differ from those of the hastingsite in the other rocks in several respects: (1) The Z direction is bluish green instead of olive green or dark green; (2) the optic angle is smaller; and (3) dispersion is stronger with $r<v$ instead of $r>v$. The blue-green color and the strong dispersion of this mineral, together with its close association with riebeckite, suggests the presence of abundant soda and ferric iron. However, it is believed that the above optical properties justify classifying this mineral as hastingsite.

The riebeckite has two modes of occurrence. Most of it forms large, irregular interstitial crystals, 1 to 3 millimeters in diameter, between subhedral grains of microperthite. These crystals are commonly spongy with poikilitic inclusions of fluorite and magnetite. Riebeckite also occurs in groups of slender, radiating fibers. Its close association with hastingsite in some sections, suggests late hydrothermal alteration of hastingsite by solutions rich in soda and iron.

The optical properties of the riebeckite from the hastingsite-riebeckite granite

TABLE 3.—*Optical properties of amphiboles from rocks of Pliny region*

	1	2	3	4	5	6	7	8
Indices α β γ	1.670* 1.683 1.689	1.678 1.693 1.699	1.688 1.699 1.704	1.706 1.725 1.729	1.689 1.710 1.712	1.689 1.704 1.711	1.698 1.719 1.722	1.691 1.694 1.699
Sign	Negative	Negative	Negative	Negative	Negative	Negative	Negative	Positive
2 V	65	65	ca. 35°	48°±2°	25°±5°	ca. 65°	47°±3°	Large
Dispersion	$r > v$, medium	$r > v$, weak	$r > v$, strong	$r > v$, medium	$r < v$, strong	$r > v$, medium	Medium strong	$r > v$, strong
Orientation	$Z \wedge c = 18^\circ$ $Y = b$	$Z \wedge c = 18^\circ$ $Y = b$	$Z \wedge c = 25^\circ$ $Y = b$	$Z \wedge c = 10^\circ$ $Y = b$	$Z \wedge c = 20^\circ$ $Y = b$	$Z \wedge c = 18^\circ$ $Y = b$	$Z \wedge c = 20^\circ$ $Y = b$	$X \wedge c = 3^\circ - 5^\circ$ $Z = b$
Pleochroism	X = light brown Y = olive brown Z = olive green $X < Y < Z$	X = light brown Y = olive brown Z = olive green $X < Y < Z$	X = yellow-brown Y = olive green Z = dark green $X < Y < Z$	X = light brown Y = olive brown Z = olive green $X < Y < Z$	X = light brown Y = olive brown Z = bluish green $X < Y < Z$	X = light brown Y = olive brown Z = olive green $X < Y < Z$	X = light yellow-brown Y = olive green Z = deep green $X < Y > Z$	X = deep blue-gray Y = pale yellow Z = gray $X > Z > Y$

1. Hornblende from quartz monzodiorite (specimen 87).*
2. Hornblende from quartz monzodiorite (specimen 120).
3. Hornblende from syenite in the Percy quadrangle, New Hampshire (Chapman and Williams, 1935, p. 512–513).
4. Hastingsite from hastingsite-quartz syenite (specimen 34).
5. Hastingsite from hastingsite-riebeckite granite (specimen 18).
6. Hastingsite from granite porphyry (specimen 45).
7. Hastingsite from quartz syenite in the North Conway quadrangle (Billings, 1928, p. 109).

Riebeckite from hastingsite-riebeckite granite (specimen 96).

* Specimen numbers correspond to specimen numbers in Table 1

(specimen 96) on the south slope of Mt. Cabot are given in Table 3. These properties are identical with those of the riebeckite from the riebeckite granite of the Percy quadrangle to the north (Chapman and Williams, 1935, p 512-513). A chemical analysis of the riebeckite from the Percy quadrangle is shown in Table 5.

HASTINGSITE-BIOTITE GRANITE

The hastingsite-biotite granite crops out east and southeast of Round Mountain, and is approximately 2 square miles in areal extent. The exact shape of the body is not known, but the regional distribution of outcrops suggests an elliptical stock.

The hastingsite-biotite granite is a medium-grained to fine-grained pink rock consisting essentially of feldspar and quartz mottled with hastingsite and biotite. The medium-grained phase is equigranular to subporphyritic with lathlike crystals of microperthite which show Carlsbad twins. The average grain size ranges from 1 to 3 millimeters, but a few biotite flakes are 5 millimeters across. The fine-grained phase shows the same textural features as the medium-grained phase, but the average grain size is slightly less than 1 millimeter.

Under the microscope the hastingsite-biotite granite is allotriomorphic and hypidiomorphic. Microperthite occurs as rectangular or irregular crystals in which the ratio by volume of potash feldspar to soda feldspar is approximately 3:1. Adjacent microperthite crystals are separated by narrow zones of tiny albite grains, many of which show albite twinning. The general appearance of these small grains suggests that they were introduced between the microperthite crystals during the late stages of crystallization. In fact this albite and that which is present as veins and patches in the microperthite may be genetically related Sodic oligoclase (An_{10}) occurs as well-twinned rectangular crystals many of which show distinct zoning. The quartz forms irregular, strained, interstitial masses between the feldspars.

The hastingsite is like that described previously. Its occurrence and optical properties are identical with those of the hastingsite in the hastingsite-quartz syenite (specimen 34) and further description is not necessary here.

Biotite occurs as irregular grains from 0 2 to 1 millimeter in diameter. Practically all grains are greatly altered at their borders and along cleavages to green chlorite, magnetite, and hematite, preventing accurate determinations of the refractive indices. However, the occurrence and pleochroism of the mineral suggests that it is not greatly different from the biotite in the other rocks of this series.

Accessory minerals include apatite, magnetite, zircon, sphene, and allanite. Secondary minerals, altering from biotite, are green chlorite and hematite. Modes of three specimens of hastingsite-biotite granite are shown in Table 1.

GRANITE PORPHYRY

The granite porphyry, which makes up the Crescent Range, covers about 5 square miles. It is confined entirely to this area except for a small dike which cuts the hastingsite-riebeckite granite at an elevation of 3200 feet on the southeast slope of the eastern knoll on Terrace Mountain.

The granite porphyry is a tough, resistant rock which is easily recognized by its peculiar texture. It is pink to gray and consists of phenocrysts of pink feldspar and smoky quartz in a dense, light-gray groundmass peppered with dark minerals.

TABLE 4.—*Optical properties of biotites from rocks of Pliny region*

		1	2	3	4	5
Indices	α	1.599	1.606	1.583	1.596	1.606
	β	1.665	1.671	1.653	1.656	1.671
	γ	1.665	1.672	1.653	1.656	1.672
Sign		Negative	Negative	Negative	Negative	Negative
2 V		ca. 3°	ca. 3°	ca. 4°	5°	3°
Dispersion		$r < v$, medium	$r < v$, medium	$r < v$, medium	$r < v$, weak	$r < v$, medium
Orientation		$X \wedge c = 3°$	$X \wedge c = 3°$	$X \wedge c = 3°$	$X \wedge c = 4°$	$X \wedge c = 3°$
Pleochroism		X = light yellow Y = red-brown Z = red-brown X < Y = Z	X = light yellow Y = brown Z = brown X < Y = Z	X = light yellow Y = brown Z = brown X < Y = Z	X = light brown Y = brown Z = brown X < Y = Z	X = light yellow Y = brown Z = brown X < Y = Z

1. Biotite from quartz monzodiorite (specimen 87).*
2. Biotite from quartz monzodiorite (specimen 120).
3. Biotite from pink biotite granite (specimen 116).
4. Biotite from gabbro at Tripyramid Mountain, New Hampshire (Chapman and Williams, 1935, p. 512–513).
5. Biotite from Conway granite (specimen 36).

* Specimen numbers correspond to specimen numbers in Table 1.

The smoky quartz phenocrysts, 1 to 2 millimeters in diameter, are always well rounded, a feature found only in the granite porphyry. The pink feldspar phenocrysts are square or rectangular and 1 to 3 millimeters in cross section. In most cases the groundmass is so fine-grained that individual crystals are not discernible megascopically.

Irregular or rounded fragments, 1 inch or less in diameter, were found in the granite porphyry. These are composed of very dense, dark-gray, crystalline material spotted with pink feldspar phenocrysts 2 to 3 millimeters across. They are believed to represent inclusions which were incorporated and reworked by the granite porphyry during its intrusion.

Microscopically the granite porphyry has much the appearance of an extrusive rock. The groundmass, composing 25 to 75 per cent, is a remarkably uniform fabric of crystals of orthoclase and quartz approximately 0.05 millimeter in diameter. Slightly more than half the groundmass is orthoclase, and the remainder is quartz with a few grains of sodic oligoclase (An_{10}).

In cross section many of the quartz phenocrysts suggest the bipyramid so common in lavas. Other quartz grains are well rounded, and some are corroded; many show evidence of straining. The feldspar phenocrysts are all microperthite, and, like the quartz crystals, generally evince rounding and corrosion.

Hastingsite occurs as small rounded or subhedral phenocrysts which range from 0.1 to 0.5 millimeter in diameter. It is notably spongy or poikilitic with numerous rounded inclusions of quartz and orthoclase. A sample was isolated from specimen 45, and its optical properties are shown in Table 3.

Biotite occurs in all specimens of granite porphyry as irregular phenocrysts 0.2 to 0.5 millimeter across. It is so badly altered to chlorite and hematite, however, that its optical properties could not be measured.

Accessory minerals are hedenbergite, apatite, magnetite, zircon, sphene, allanite, and fluorite. Biotite alters to green chlorite and hematite. Six modal analyses of the granite porphyry are shown in Table 1.

According to Billings (personal communication) the granite porphyry is similar to the Mt. Lafayette granite porphyry of the Franconia quadrangle, New Hampshire (Williams and Billings, 1938).

PINK BIOTITE GRANITE

The pink biotite granite forms dikes and small irregular intrusive masses in the composite ring dike immediately northeast of Jefferson. Small masses, not shown on Plate 1, were also found 1 mile north of the village of Starr King.

This granite is pinkish-gray and shows a uniform character in all outcrops. Subporphyritic varieties occur, but on the whole the rock is remarkably even grained with individual crystals averaging 1 millimeter in diameter. Some specimens show a crude alignment of feldspar laths. The essential minerals are potash feldspar, quartz, and biotite.

Thin sections show this rock to be closely similar to the Conway granite, of which it may indeed be a phase. The texture is hypidiomorphic and even grained, and adjacent grains commonly interlock. Microperthite is the chief mineral, and its ratio by volume of potash feldspar to soda feldspar is 3:1. Sodic oligoclase (An_{15}) and interstitial quartz are the other essential felsic constituents.

TABLE 5.—*Chemical composition of minerals of White Mountain magma series*

	1	2	3	4	5	6	7
SiO_2	37 74	35 88	33.48	47.58	46.56	37.40	46.98
TiO_2	5.22	4 17	2 94	37	1.83	3.20	1 49
Al_2O_3	16 31	14 96	13.64	1 16	2 48	12 34	1 29
Fe_2O_3	none	2.33	8.00	2.60	6 24	4 16	11 93
FeO	15.52	20.88	23 54	24 21	27.27	25.84	23 38
MnO	.06	.18	1 02	59	1.08	1.24	.24
MgO	14 23	10.04	4 97	3 34	1.35	2 20	.13
CaO	04	.12	.56	18 80	6 15	9 72	1 91
Na_2O	.44	36	53	.47	4.17	1 80	8 90
K_2O	8 88	9.20	7 80	21	1 37	1.36	2 74
H_2O+	1.17	.80	2 63	.34	1 27	tr	1 10
H_2O-					nd	.60	none
F	.45	1 58	.95	....	......	....	....
	100 06	100.50	100.06				
Less O for F	.19	66	.40				
Total	99 87	99 84	99.66	99 67	99 77	99 86	100 09

1. Biotite from gabbro (7052)*, Black Cascade, Tripyramid Mountain.
2. Biotite from monzodiorite (609), one-third mile northwest of Ames Station, Belknap Mountains.
3. Biotite from granite (19287), granite quarry 1 mile south of Beech Hill, Percy quadrangle.

* Numbers in parentheses refer to Harvard University rock collections. F. A. Gonyer, analyst of all except specimen 6. All analyses except number 6 quoted from Chapman and Williams (1935, p 512).

4. Hedenbergite from syenite (19258), east slope of knoll south of Burnside Brook at elevation 2160 feet, Percy quadrangle.
5. Hornblende from syenite (19247), in Moore Brook at elevation 2000 feet, Percy quadrangle.
6. Hastingsite from quartz syenite (18299), Jackson Falls, Jackson, North Conway quadrangle, (Billings, 1928, p. 110).
7. Riebeckite from granite (19274), one-quarter mile northeast of small swamp on Mill Mountain, Percy quadrangle.

Biotite is the only primary ferromagnesian mineral. In some specimens clean, irregular grains measure 0.2 to 1 millimeter across. Pleochroic halos surrounding zircon inclusions are common. In other specimens, all the biotite has altered to chlorite. The optical properties of the biotite from specimen 116 resemble those of a biotite from a gabbro of the White Mountain magma series from Tripyramid Mountain, New Hampshire (Chapman and Williams, 1935, p. 512-513). These are compared in Table 4.

The birefringence of the biotite from the pink biotite granite is higher than that of the biotite from the Tripyramid Mountain gabbro. In fact it is higher than that of any biotite of the White Mountain magma series determined to date. This may be due to abundant ferric iron and/or titanium (Winchell, 1933, p. 273). The chemical analysis of the biotite from the gabbro of Tripyramid Mountain (Chapman and Williams, 1935, p. 512) is shown in Table 5. In spite of somewhat similar indices, the biotite from the pink biotite granite may be much richer in iron than the biotite from the Tripyramid Mountain gabbro.

Accessory minerals of the pink biotite granite are apatite, magnetite, zircon, sphene, and allanite. Secondary chlorite from biotite is common. Two modes of the pink biotite granite are shown in Table 1.

CONWAY GRANITE

The Conway granite forms numerous, small, probably more or less circular stocks (Pl. 1). It also occurs at the very northern edge of the area on the southern margin of a large stock which extends south from the Percy quadrangle. The total area of the Conway granite exposed in the Pliny region is about 4 square miles.

This granite, an important member of the White Mountain magma series, is a granular rock, composed principally of pink potash feldspar, smoky quartz, and biotite. Each of these minerals collects into aggregates. The rock is typically pink and weathers white or brownish. Its texture varies widely so that in some other areas it has been divided into several textural phases (Billings, 1928, p. 118-124; Chapman, 1935, p. 428-430). In the Pliny region no attempt was made to map variations because the bodies are small and isolated and the outcrops are poor.

In the Pliny region the Conway granite is generally medium-grained with individual crystals ranging from 1 to 5 millimeters in diameter, but some specimens are fine-grained. A porphyritic phase, extending into the Pliny region from the north, is exposed in a small quarry along the unimproved road 1 mile west-southwest of York Pond. The groundmass is similar in texture and grain size to that of the medium-grained granite, but the lathlike phenocrysts of microperthite attain a diameter of nearly 15 millimeters. Most of these phenocrysts show Carlsbad twinning.

Microscopically the Conway granite is hypidiomorphic-granular and consists essentially of microperthite, slightly strained quartz, twinned and zoned oligoclase (An_{15}), and brown biotite. Accessory minerals are apatite, magnetite, zircon, sphene, and allanite. Other accessories have been reported from other areas. Chlorite occurs as an alteration from biotite.

The nature and occurrence of the biotite in this rock are the same as for the biotite of the pink biotite granite, but the indices of refraction are much higher.

The optical properties of the biotite in a specimen (36) of Conway granite from the small body 1⅛ miles west-northwest of Mt. Starr King are shown in Table 4. They match those measured on biotite from the Conway granite in the Percy quadrangle, New Hampshire (Chapman and Williams, 1935, p. 512-513). The chemical analysis of the biotite from the Percy quadrangle is shown in Table 5.

AGE RELATIONS OF ROCKS OF THE WHITE MOUNTAIN MAGMA SERIES

The relative ages of the rock members of the White Mountain magma series in the Pliny region are based upon: (1) relations as determined at contacts between these rock types in the Pliny region, and (2) correlation of these rocks with their equivalents in other regions where the relative ages are known.

The quartz monzodiorite and quartz monzonite are similar petrographically and intergradational, and they are, therefore, probably essentially contemporaneous. Fortunately, there is considerable evidence in the Pliny region as to the relative ages of these two rock types.

In Garland Brook, where the quartz monzodiorite and quartz monzonite disappear beneath the drift, they are cut by dikes and stringers of white, medium-grained biotite granite. The former are, therefore, older than the granite which has been identified as Conway granite, the youngest plutonic rock in the region. Northeast of Jefferson, and in the "shatter zone" of the composite ring dike, irregular inclusions of quartz monzodiorite are abundant in the pink biotite granite. Thus the quartz monzodiorite is older than the pink biotite granite. Similarly, the quartz monzodiorite is older than the hastingsite-riebeckite granite because on the southeast slope of the eastern knoll on Terrace Mountain, at an elevation of 3200 feet, a large inclusion of quartz monzodiorite was found in the hastingsite-riebeckite granite.

At the north end of the ring dike of quartz monzodiorite and quartz monzonite, northwest of Mt. Starr King, a small lens of hastingsite-quartz syenite intrudes the eastern boundary between the dike and the country rock. Tongues of medium-grained hastingsite-quartz syenite cut the dike rock which is here a quartz monzonite, and blocks of quartz monzonite are included in the hastingsite-quartz syenite, indicating clearly the age relations. Similar contact relations were observed farther south and on the west side of the ring dike of quartz monzodiorite and quartz monzonite where it comes in contact with the hastingsite-quartz syenite in the northern part of the composite ring dike.

Nothing more can be said about the age relations of the quartz monzodiorite and quartz monzonite to the other rocks, from data gathered here. However, studies in other areas, indicate that the quartz monzodiorite and quartz monzonite are the oldest rocks of the White Mountain magma series in the Pliny region.

The relative age of the granite porphyry composing the Crescent Range ring dike is somewhat uncertain since it is entirely isolated from other intrusions of the White Mountain magma series. However, a small dike of granite porphyry, very similar to that of the Crescent Range, cuts the hastingsite-riebeckite granite at an elevation of about 3200 feet on the southeast slope of the eastern knoll of Terrace Mountain. The similarity and rather unusual nature of the two granite porphyries suggest that they are contemporaneous. Since the granite porphyry cuts the hastingsite-riebeckite granite it must be younger.

Since the hastingsite-riebeckite granite stock apparently cuts off the ring dike of hastingsite-quartz syenite it must be younger. On the other hand, the hastingsite-riebeckite granite is older than the Conway biotite granite because it is cut by three small intrusions of Conway granite in the Pliny region, and by one large stock in the Percy quadrangle (Pl. 1).

The hastingsite-quartz syenite is older than the Conway biotite granite, because it has been intruded by an elliptical stock of Conway granite south of Mt. Starr King, and by a similar but smaller stock due east of Round Mountain. This view is substantiated by the fact that in the Percy region (Chapman, 1935, p. 415-416) the Conway granite has been shown to be younger than the hornblende syenite (similar to the hastingsite-quartz syenite of the Pliny region).

Regional mapping suggests that the large stock of hastingsite-biotite granite southeast of Round Mountain has cut out the hastingsite-quartz syenite ring dike, and that the hastingsite-biotite granite is thus younger than the hastingsite-quartz syenite.

The hastingsite-biotite granite resembles the Conway biotite granite petrographically and the two are probably nearly contemporaneous. The pink biotite granite is doubtless a phase of the Conway biotite granite. As in other regions, the Conway biotite granite in the Pliny region seems to be the youngest plutonic rock of the White Mountain magma series.

In summary, the intrusive rocks of the White Mountain magma series occurring in the Pliny region, excluding complementary dikes, are arranged chronologically as follows, oldest at the bottom:

Conway granite
Pink biotite granite[3]
Granite porphyry[3]
Hastingsite-biotite granite[3]
Hastingsite-riebeckite granite
Hastingsite-quartz syenite
Quartz monzodiorite and quartz monzonite

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[3] The exact position in the sequence of these rocks is not certain.

TABLE 6.—*Structural units of Pliny region, New Hampshire* *

Structural unit	Shape	Dimensions	Area	Rock type
Conway granite stocks	Circular to elliptical	Diameter 1/4 to 1 mile. For large stock near York Pond *see* Plate 1	1/15 to 1 sq. mile. For large stock near York Pond *see* Plate 1	Pink, generally medium-grained biotite granite
Composite ring dike (southern portion)	Arcuate	Radius of curvature 3 1/2 miles	1 sq. mile	Pink to gray, even-grained to subporphyritic biotite granite
Crescent Range ring dike	Arcuate	Radius of curvature 2 1/2 miles	5 sq. miles	Pink to gray granite porphyry
Hastingsite-biotite granite stock	Probably roughly elliptical	Long diameter 2 3/4 miles. Short diameter 1 1/2 miles	2 sq. miles	Pink, medium-grained to fine-grained hastingsite-biotite granite
Hastingsite-riebeckite granite stock	Roughly elliptical	Long diameter 5 miles. Short diameter 3 miles	Complete stock 10 sq. miles. Portion in Pliny region 6 1/2 sq. miles	Medium-grained hastingsite-riebeckite granite. Fresh specimens blue green
Composite ring dike (northern portion)	Arcuate	Radius of curvature 3 1/2 miles	1/5 sq. miles	Medium-grained hastingsite-quartz syenite. Fresh specimens blue green
Hastingsite-quartz syenite stock	Unknown	Unknown	Unknown	Medium-grained hastingsite-quartz syenite. Fresh specimens blue green
Hastingsite-quartz syenite ring dike	Asymmetrically arcuate	Radius of curvature 2 1/2 miles	5 to 6 sq. miles	Medium-grained hastingsite-quartz syenite. Fresh specimens blue green
Ring dike of quartz monzodiorite and quartz monzonite	Arcuate	Radius of curvature 3 miles	1 sq. mile	Dark-gray, medium-grained quartz monzodiorite and quartz monzonite
Body of quartz monzodiorite and quartz monzonite	Circular or arcuate	Average diameter 3 1/2 miles	At least 4 1/2 sq. miles	Dark-gray, medium-grained quartz monzodiorite and quartz monzonite

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* Arranged chronologically, as nearly as possible; oldest at bottom.

STRUCTURE OF THE RING DIKE COMPLEX

DESCRIPTION OF STRUCTURAL UNITS

General statement.—The structural units of the Pliny region are ring dikes and stocks, intrusive into older crystalline rocks. Most of the ring dikes and stocks are concentric about a center in the interior of the Pliny Range, although the center of the Crescent Range ring dike is 3 miles to the east-southeast. The ring dikes form only arcs or crescents, with radii of from 2½ to 3½ miles. The largest stock has an average diameter of 3½ miles, whereas the smallest is about ¼ mile across. It is possible to arrange these structures in a definite chronological order, as in Table 6, since the relative ages of the rocks, which belong to the White Mountain Magma series, have been determined.

In the following pages each structural unit will be considered in detail. For the sake of simplicity and convenience these will be discussed according to type rather than according to age, i.e. the ring dikes will be considered first and then the stocks.

Body of quartz monzodiorite and quartz monzonite.—This body underlies the circular basin north of the Pliny Range (Pl. 1), at least in part, and is either an arcuate ring dike or a circular stock averaging approximately 3½ miles in diameter. It is apparently cut off on the north by the hastingsite-riebeckite granite stock. The best exposures of quartz monzodiorite and quartz monzonite occur on Round Mountain and on and around the two peaks to the south.

The outer contact of this body, although not exposed, has been located rather accurately by regional mapping. Whether it is vertical or inclined is not known. Unfortunately, there are no structural features such as foliation or lineation to indicate the attitude of the contact.

Ring dike of quartz monzodiorite and quartz monzonite.—This long, slender, arcuate dike 1½ miles west of Mt. Starr King and north of Jefferson trends somewhat west of north and has a radius of about 3 miles. It is 4 miles long, has a maximum thickness of a third of a mile, and tapers on both ends. Its areal extent is about 1 square mile. The quartz monzodiorite and quartz monzonite resemble that of the last body described. Outcrops are fairly abundant, particularly along the southwesterly trending ridges due west of Mt. Starr King.

The boundary, although not exposed, has been fairly accurately located except at the southeastern end where outcrops are scarce. It is probably steep since it trends undeflected across rather rugged topography. The structure of the country rock (syenite of the Oliverian magma series) was not greatly affected by this intrusion, for the foliation near the dike conforms in general to the regional trend.

Hastingsite-quartz syenite ring dike.—This ring structure, covering 5 or 6 square miles, is the largest in the Pliny region. It underlies most of the Pliny Range and is responsible for the great height and crescentic shape of the latter. Inside the arc the dike is bounded by the body of quartz monzodiorite and quartz monzonite and outside the arc by syenite of the Oliverian magma series. In ground plan the dike is an asymmetrical crescent or a portion of a ring whose major axis is 5 miles long and trends north-northeast. The minor axis of the ring is 4¼ miles long and trends east-southeast. Originally the dike may have been a complete ring, now cut off on the north by the hastingsite-riebeckite granite stock.

Despite the rugged topography, outcrops are not plentiful. The best exposures occur at the northwest end of the dike (north of hill, elevation 3343 feet) and along the east slope of the Pliny Range southeast of Round Mountain. The tough, massive hastingsite-quartz syenite forms rugged cliffs, up to 50 feet or more in height.

The contact between the hastingsite-quartz syenite and the surrounding country rock is not exposed, and where outcrops are lacking, it is mapped with a dashed line. Where the contact has been fairly definitely located—*i.e.*, in the region northwest of Mt. Starr King and in the vicinity of Round Mountain—it crosses rugged topography without deflection, suggesting that it is steep.

Composite ring dike.—The composite ring dike is a long, very slender, somewhat irregular series of intrusions west, southwest, and south of the Pliny Range. Its arcuate trend closely parallels that of the hastingsite-quartz syenite ring dike. It is 7 miles long and its thickness varies greatly. From the vicinity of Jefferson northward it is 500 to 700 feet thick; southeast of Jefferson, however, it becomes a zone of intrusion, and attains a maximum thickness of nearly 1500 feet.

The best exposures occur along the westerly flowing streams and on the intervening ridges north of Jefferson. Southeast of Jefferson, outcrops are generally poor.

The trend of this body seems to be more or less independent of the structure of the surrounding rock (Pl. 1). Northwest of Jefferson, the foliation of the country rock trends northeast and dips northwest. Southeast of Jefferson, however, the foliation strikes northwesterly and dips northeast. Whether these structural complexities were caused by the emplacement of the composite dike or by an earlier deformation cannot be determined.

This ring dike is called a composite dike because it consists of two distinct portions. The portion north of a line extending due west from the summit of Mt. Starr King is more or less uniform in thickness and is composed of medium-grained hastingsite-quartz syenite. It apparently emplaced itself along the western boundary of the ring dike of quartz monzodiorite and quartz monzonite. Half a mile south of the north end of the composite dike, and on its east side, tongues and stringers of hastingsite-quartz syenite cut and include fragments of the quartz monzodiorite and quartz monzonite. Many of these inclusions show signs of partial assimilation. Inclusions of black, fine-grained rock, which possibly belongs to the Lost Nation group, also occur here. On the west side of the composite dike, stringers of hastingsite-quartz syenite intrude coarse, pink syenite of the Oliverian magma series. No data are available on the attitude of the dike walls.

South of a line extending due west from the summit of Mt. Starr King the composite ring dike becomes a zone of intrusion. Accordingly, its thickness varies and its boundaries are not sharp. A screen of country rock intervenes between this zone of intrusion and the ring dike of quartz monzodiorite and quartz monzonite. The country rock in the zone is coarse, pink, porphyritic syenite of the Oliverian magma series with some inclusions of black, fine-grained schist and amphibolite. It is cut by numerous, scattered, irregular or dikelike intrusions of medium-grained to fine-grained, pink biotite granite from a few inches up to 10 feet thick. Numerous, irregular, blocklike inclusions of syenite of the Oliverian magma series, cut by small stringers of pink biotite granite, are common in the granite intrusions.

In the shatter zone, northeast of Jefferson, irregular inclusions of quartz monzodiorite occur in pink biotite granite dikes. Many of these inclusions are penetrated by slender biotite granite stringers. One particular granite dike, filled with inclusions, has sharp, regular walls indicating that the inclusions were brought up at least some distance as the granite moved into place. A few dikelike intrusions of quartz monzodiorite cut the Oliverian syenite. One dike, 30 feet long and 1 foot thick, is cut and displaced along shear fractures observable in both the dike and the country rock. The intrusion of pink biotite granite may account for this displacement.

Fine-grained, dark trap dikes cut the syenite of the Oliverian magma series in this zone. Most of these are probably older than the White Mountain magma series since they are slightly metamorphosed. In two localities these trap dikes are intruded by the pink biotite granite.

The southeastern part of the composite ring dike appears to be a shatter zone into which pink biotite granite has been intruded as dikes and irregular masses. Modell (1936, p. 1906) found a similar phenomenon, but on a much smaller scale, in the ring dike of Albany quartz syenite in the Belknap Mountains.

The quantity of quartz monzodiorite in the shatter zone is so small and the distribution so local that it is impossible to prove whether the rock was intruded before or after shattering. However, the abundant, irregular fragments of quartz monzodiorite believed to have been brought up from depth by the pink biotite granite, suggest that the quartz monzodiorite antedated the shattering. In other words, the quartz monzodiorite apparently was not intruded into the shatter zone, and is not, therefore, a part of the composite ring dike.

As the pink biotite granite is probably younger than the hastingsite-quartz syenite the ring dike is truly composite. The northern hastingsite-quartz syenite portion of the dike was probably contemporaneous with the large hastingsite-quartz syenite ring dike of the Pliny Range. The southeastern part is younger and consists of a shatter zone containing numerous pink biotite granite dikes.

Crescent Range ring dike.—This body of pink granite porphyry forms the backbone of the Crescent Range. Its arcuate shape, regular boundaries, and remarkable symmetry are strong indications that it was emplaced along either a circular, elliptical, or arcuate fracture, although the fracture has not been observed beyond the limits of the intrusion. In its northern portion the dike and mountain range are coextensive, but toward the south, in the vicinity of Mt. Crescent, the dike swings westward, and the mountain ridge continues to the southwest for nearly 5 miles (Pl. 1). Along the arc, the dike measures nearly 8 miles, and the maximum width, just south of Black Crescent Mountain, is 1 mile. The dike becomes progressively thinner until it finally pinches out at both ends. It is not known whether the dike was emplaced along a circular or elliptical fracture, but if we assume the former, the radius of curvature, 2½ miles, is about equal to that of the hastingsite-quartz syenite ring dike. However, the areal extent (about 5 square miles) is slightly less than that of the hastingsite-quartz syenite dike.

The Crescent Range ring dike is probably the best-exposed structural unit in the region. Outcrops are particularly abundant in the vicinity of Mt. Crescent and along the easterly flowing streams north and south of Black Crescent Mountain. Near the horns of the crescent, outcrops are scattered. On the whole, exposures are better on the east side of the range than on the west side where glacial drift is thick.

A sharp contact between the ring dike and the pink gneiss of the Oliverian magma series is exposed at an elevation of 2540 feet in the southeasterly flowing stream which heads half a mile northeast of Mt. Crescent. The contact is sharp and distinct, and although the dip could not be measured accurately, it is apparently very steep. The contact has not been observed elsewhere, but it has been located within a few hundred feet at several places, both in deep valleys and on some high ridges. When these points are connected, smooth, curved boundaries are produced, without deflection by topography, implying that the walls of the dike are very steep if not vertical. Specifically, the contact has been located within about 100 feet horizontally on the summit of Mt. Crescent. Its altitude is 3230 feet. In the stream valley half a mile northeast of Mt. Crescent the contact is exposed at an altitude of 2540 feet. Thus there is a vertical difference of 690 feet within a horizontal distance of half a mile, and yet both points lie on the smooth curve determined by several other such points.

Locally, near the northeastern boundary of the ring dike, the foliation of the country rock seems to have been diverted from its general northeasterly trend. When considered regionally, however, the Crescent Range ring dike cuts directly across the structure of the country rock (Pl. 1).

At an elevation of 3200 feet in the hastingsite-riebeckite granite on the southeast slope of the eastern knoll on Terrace Mountain is a small dike of granite porphyry identical with that of the Crescent Range. This dike, although small, lies on what might be considered the projection of the north end of the Crescent Range ring dike. It may represent an intrusion along the same ring fracture. In any event, since the small dike on Terrace Mountain is similar petrographically to the Crescent Range ring dike, and since the rock of which both are composed is a rather unusual type, it appears that both were intruded contemporaneously.

Hastingsite-quartz syenite stock.—One mile southwest of Pliny Mountain is a large outcrop of medium-grained to fine-grained hastingsite-quartz syenite. There is no information as to the size or shape of the mass; consequently, it is outlined on the geological map (Pl. 1) with dashes. As this body petrographically resembles the hastingsite-quartz syenite ring dike of the Pliny Range, the two were probably more or less contemporaneous.

Hastingsite-riebeckite granite stock.—The hastingsite-riebeckite granite stock lies mainly in the Mt. Washington quadrangle, but the mass extends north for about 1 mile into the Percy quadrangle. The outline of the body is roughly elliptical with the long axis, nearly 5 miles, trending northwest-southeast. The short axis is approximately 3 miles long. The area of the stock (excepting inclusions and younger intrusions) is about 10 square miles, but the portion in the Pliny region covers about 6½ square miles.

In the Percy quadrangle, the stock is not well exposed, but outcrops may be seen near the summit of Mt. Cabot and on the ridge directly west between elevations 2500 feet and 3000 feet. In the Pliny region, exposures are excellent along the crest and upper slopes of Terrace Mountain.

Since the contacts of the hastingsite-riebeckite granite stock are not exposed, their dips are not known. The eastern boundary has been rather accurately located, but because of poor exposures the western boundary can be established only in a general way. On the south the contact is covered by glacial drift.

In 1935 when R. W. Chapman published on the ring dike complex of the Percy quadrangle, little was known about the geology of the Pliny region, and this body of hastingsite-riebeckite granite, was believed to be a thick ring dike. However, present studies show the body to be a stock, and it was, therefore, necessary to alter somewhat those boundaries which had been drawn in the Percy region.

The hastingsite-riebeckite granite stock contains abundant quartzite inclusions, many of which may be seen in the well-exposed granite along the crest of Terrace Mountain. Others occur on the flanks of Terrace Mountain and on the south slope of Mt. Cabot. The outlines of these inclusions are probably irregular, but only limited portions of their boundaries are visible. They range in cross section from a few hundred feet to nearly 3000 feet, the largest one lying less than 1 mile northeast of the northernmost peak of Terrace Mountain. The inclusions consist of dense, dark quartzite of the Albee formation, like the country rock to the west. Most of the blocks are well stratified, and show random strike and dip.

Hastingsite-biotite granite stock.—This body of pink, medium- to fine-grained hastingsite-biotite granite lies east and southeast of Round Mountain, adjacent to the ring dike of hastingsite-quartz syenite. Due to the paucity of outcrops, the exact shape of the body is unknown, and its boundaries, therefore, are mapped with a dashed line. The body covers an area of about 2 square miles. Its north-south axis is 2¾ miles, and its east-west axis is 1½ miles. The distribution of outcrops at least suggests that the hastingsite-biotite granite stock has cut into the hastingsite-quartz syenite ring dike and that the former is, therefore, younger.

Conway granite stocks.—In the Pliny region the Conway granite is represented by numerous small stocks, six of which are mapped (Pl. 1). They are all roughly circular in plan, although in many cases boundaries could not be located accurately. The largest of these stocks lies due south of Mt. Starr King where it intrudes the ring dike of hastingsite-quartz syenite. A smaller one cuts the hastingsite-quartz syenite ring dike due east of Round Mountain, and a third occurs in the country rock (pink syenite of the Oliverian magma series) 1⅓ miles west-northwest of Mt. Starr King. Three more small stocks intrude the hastingsite-riebeckite granite on the flanks of Terrace Mountain. Many others, too small to map, were found cutting the older rocks in several places. The abundance and distribution of these bodies suggest that they represent cupolas rising from a much larger intrusive mass below.

The southern portion of a large stock of Conway biotite granite lies along the northern edge of the Mt. Washington quadrangle just east of the hastingsite-riebeckite granite stock. It is well exposed in a small quarry along the unimproved road 1 mile west-southwest of York Pond. Although its southern boundary is mapped with a dashed line, its location is believed to be fairly accurate since it is extrapolated from detailed mapping farther north. This large stock apparently cuts out part of the hastingsite-riebeckite granite stock east of Mt. Cabot.

RELATION OF THE PLINY REGION TO OTHER RING DIKE AREAS

General statement.—Before attempting to determine the origin of the structural units of the Pliny region, it will be helpful to consider the relation of this area to other New England districts in which ring dikes have been described: the Ossipee Mountains, New Hampshire (Kingsley, 1931), the Percy region, New Hampshire (R. W. Chapman, 1935), the Belknap Mountains, New Hampshire (Modell, 1936), the Franconia quadrangle, New Hampshire (Williams and Billings, 1938), the Mt. Chocorua quadrangle, New Hampshire (Smith, *et al.*, 1939), and Ascutney Mountain, Vermont (Chapman and Chapman, 1940).

Proximity to the Percy region.—The Percy region bounds the Pliny region on the north. The topography, rock types, and structural features of these two areas are very similar; the two regions are part of one large intrusive complex which covers an area of nearly 200 square miles.

Similarity of structural units.—Some of the structural units of the Pliny region resemble those in other New England areas. For example, although the ring dike of the Crescent Range is much longer, covers a greater area, and is composed of granite porphyry rather than syenite porphyry, it has the same crescentic shape and the same radius of curvature as the Cape Horn ring dike of the Percy complex. Field relations indicate that the walls of both dikes are very steep: the Cape Horn dike is believed to dip about 75°, and the Crescent Range dike may dip even more steeply.

A second comparison may be made between the hastingsite-quartz syenite ring dike of the Pliny Range and the Albany quartz syenite ring dike of the Belknap Mountains (Fig. 1). Both are asymmetrical crescents of quartz syenite, and the radius of curvature of both is the same. The dike of the Belknap Mountains, however, is somewhat more slender and less regular.

In the third place, the composite ring dike of the Pliny region is similar to the Rattlesnake composite ring dike of the Belknap Mountains. Both are discontinuous and both are composed of several kinds of rock. The composite dike of the Belknap Mountains is considerably thicker however.

Finally, the role of the Conway granite as stocklike bodies is also common in the other ring dike areas of New England, but small ring dikes of it are found in the Belknap Mountains and in the Percy complex. The stocks in the Pliny region are smaller and more numerous than those in other areas.

Centers of intrusion.—The ringlike intrusions of the Pliny region are grouped about two separate centers. One lies in the midst of the Pliny Range about 2 miles northeast of Mt. Starr King, and it is about this center that the older ring dikes are arranged. The granite porphyry ring dike, which is probably the youngest in the region, has a center of intrusion 3 miles northeast of Pliny Mountain. The granitic stocks are distributed irregularly about the earlier center of intrusion. A similar arrangement of ring structures and associated stocks about more than one center of intrusion has been demonstrated in the Belknap Mountains, the Percy region, and at Ascutney Mountain, Vermont (Fig. 1).

Rock types.—In the Pliny region, as in the other New England ring dike areas, the intrusions belong to the White Mountain magma series. The older structural units are composed of the more basic rocks, whereas the later bodies are made of siliceous varieties. Inasmuch as the stocks are generally the latest, they are composed principally of granite. Apparently the rocks of the White Mountain magma series have followed the normal order of differentiation, and consequently the rock type composing a particular body depends upon the relative age of that body. In the case of composite ring dikes, there has been more than one period of intrusion along the same fracture.

INTRUSION OF THE RING DIKES

General theory.—Due to a lack of good rock exposures, information on the origin of the ring structures of the Pliny region is incomplete. Although there is no positive evidence of the sinking of large, cylindrical blocks, the similarity of the ring structures here to those in other New England areas where the origin has been fairly definitely established, suggests that cauldron subsidence or ring-fracture stoping has been the method of origin of the ring dikes and perhaps some of the stocks. This theory has been discussed so much recently in the literature that only a résumé of the essential features will be given here.

The concept of cauldron subsidence or ring-fracture stoping was developed by E. M. Anderson (Clough, *et al.*, 1909, p. 11-12). He assumes a paraboloidal magma chamber in the crust at a depth of several miles. As long as the roof and magma chamber remain under the same hydrostatic pressure no fractures will form, but if the magmatic pressure is reduced, a tension is set up. This results in a number of steeply dipping fractures that slope outward. The sinking of the central block into the magma chamber and the intrusion of magma along these fractures result in ring dikes. During the sinking of the central block a portion of the ring fracture may become closed up in one manner or another so that the intrusion of magma along it is impossible. Thus a partial or incomplete ring dike, having an arcuate or crescentic shape, might form. Various other modes of formation of such structures have been discussed in some detail by Modell (1936, p. 1925-1927).

The theory outlined above is general and does not explain in detail the mechanism by which the ring dikes emplace themselves. Fortunately, the Pliny region offers some evidence on this problem.

Evidence offered by the composite ring dike.—The northern portion of the composite ring dike is the older and is composed of hastingsite-quartz syenite, whereas the southern part is a more or less irregular zone with intrusions of pink biotite granite. The dike is not a solid intrusion which has emplaced itself between the two walls of a sharp fracture, but instead, it is a zone of intense fracturing into which many small dikelike intrusions have worked their way. The writer believes that the body originated somewhat as follows.

Tensional stresses in the region formed a curved fracture which now marks the general position of the dike. An isolated, cylindrical block produced by the curved fracture, settled downward into the magma chamber. As the two walls moved past one another, they shattered and crushed the rock on both sides of the break. Possibly this displacement caused considerable deformation of the earlier structures in the surrounding rock. Spreading of the fracture walls in the late stages of the movement produced a zone of porous rubble which favored magmatic intrusion. In the northern portion of this porous zone hastingsite-quartz syenite magma rose, engulfing the fractured blocks, until a more or less solid dike was formed. Quite likely piecemeal stoping was important in making room for the rising magma. A somewhat similar mechanism has been suggested by Williams and Billings (1938) in the Franconia quadrangle.

Later, and perhaps accompanied by further displacement of the walls, intrusion was initiated in the southern portion of the composite dike. Pink biotite granite rose as a network of small, irregular dikes, filling fractures and spaces between irregular blocks.

Evidence offered by the Crescent Range ring dike.—The Crescent Range ring dike differs from the composite ring dike in several ways: (1) It is a symmetrical crescent; (2) its boundaries are regular and smoothly curving; (3) it is a solid dike which lacks inclusions of any great size (a few small inclusions are discussed in the petrographic description); and (4) it is composed of granite porphyry which has corroded phenocrysts in a very dense groundmass.

It is evident from these comparisons that the granite porphyry ring dike had a quite different origin. Its symmetrical, crescent shape and its regular, smoothly curving boundaries indicate clearly that it rose en masse along a sharp, curving fracture. The sharp, regular boundaries and the lack of abundant inclusions suggest that stoping of small blocks was unimportant. In contrast with the other dikes of the area, the Crescent Range ring dike must have cooled quickly since the groundmass of the granite porphyry is dense. However, the large, corroded phenocrysts of orthoclase and quartz suggest that the magma must have been cooling slowly at some deep level prior to its final injection.

The genesis of the Crescent Range ring dike seems to have been somewhat as follows: somewhere beneath the present surface lay a body of granitic magma which was cooling slowly. Early formed crystals of orthoclase and quartz were being resorbed and corroded by the cooling melt. Apparently resorption and corrosion took place at depth rather than in the dike, since cooling of the magma was very rapid after the dike was emplaced. Then a steeply dipping, circular or slightly elliptical fracture developed across the structure of the country rock, as a result of tensional stresses. The cylindrical or domical block thus formed sank and crowded westward so that a crescentic opening instead of an annular opening was produced. Into this opening the granitic magma was injected and suddenly cooled to produce a symmetrical, crescentic ring dike of granite porphyry. The original fracture may have extended as far west as Terrace Mountain because a small dike of granite porphyry was found at an elevation of 3200 feet on the southeastern slope of the eastern knoll.

The crescent-shaped Cape Horn ring dike in the Percy region is composed of dense syenite porphyry. Its history is probably essentially the same as that of the Crescent Range ring dike. According to Billings (personal communication) the ring dike of granite porphyry forming the Franconia Range is also analogous.

Evidence offered by the other ring dikes.—The ring dike of quartz monzodiorite and quartz monzonite and the hastingsite-quartz syenite ring dike have fairly regular, broadly sweeping boundaries, and neither contains inclusions. This suggests that they were intruded en masse along sharp fractures, and that piecemeal stoping of small blocks was unimportant. Both dikes, however, are composed of medium-grained rock which implies that cooling was relatively slow. These two bodies are typical of most of the ring dikes of northern New England.

INTRUSION OF THE STOCKS

The large Conway granite stock extending south from the Percy region, the large hastingsite-riebeckite granite stock of Terrace Mountain, and perhaps even the hastingsite-biotite granite stock southeast of Round Mountain can all be explained by ring-fracture stoping if it be assumed that in each case the central block has dropped below the present erosion surface. However, the smaller stocks composed of Conway granite may represent small cupolas which have risen from a large granite mass below, by piecemeal stoping of the country rock.

ORIGIN OF THE ROCK TYPES

The rocks of the Pliny region are believed to have originated in a large reservoir of basaltic magma at some depth. Inasmuch as they exhibit a definite sequence from basic to acid types, fractional crystallization is thought to have been the dominant process of formation. At various stages during differentiation the reservoir was tapped by fractures extending toward the surface, and the magma rose along these breaks to form the ring dikes and stocks. Late granitic types are particularly abundant in the Pliny region, however, and it is difficult to account for large quantities of silica by fractional crystallization alone. Possibly in the late stages of differentiation, large masses of assimilated siliceous country rock helped to form abundant granitic magma.

The concept outlined above is the same as that proposed some years ago to explain the White Mountain magma series as a whole (Chapman and Williams, 1935).

SUMMARY AND CONCLUSIONS

The conclusions drawn from the present study are summarized as follows: (1) The Pliny region is similar to several other New England areas in that it has ring dike and stock intrusions; (2) the intrusive bodies are grouped about two distinct centers; (3) the ring dikes are probably all arcs or partial rings; (4) for the most part the stocks are younger than the ring dikes and they cut across the latter; (5) cauldron subsidence or ring-fracture stoping (the settling of large, cylindrical or domical blocks along curved fractures) is believed to be the best explanation of the intrusion of all of the ring dikes and some of the stocks; (6) the composite ring dike near Jefferson resulted from the injection of magma into a highly fractured, arcuate zone; (7) the ring dike of the Crescent Range, however, was formed by the injection en masse of granite porphyry along a smooth, sharp, curved fracture; (8) all intrusive bodies are composed of differentiates of the White Mountain magma series and are believed to have originated chiefly through fractional crystallization of a basaltic magma somewhere at depth; and (9) the presence of many granitic types among these intrusions, however, implies that assimilation of siliceous country rock was also important in producing the late granitic differentiates.

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UNUSUAL CURRENTS IN GLACIAL LAKE MISSOULA, MONTANA

BY J. T. PARDEE

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ABSTRACT

The area submerged by Glacial Lake Missoula includes several intermontane basins and constricted interconnecting valleys or "narrows" that drain to a single outlet, the Clark Fork River. A sudden failure of the ice dam that blocked this valley, near the Idaho-Montana State line, caused unusually large and rapid currents through the narrows and wind gaps in the partly submerged rim of Camas Prairie basin. Evidences of such currents include commensurate, but otherwise ordinary, effects of streams confined to rocky channels and the unique giant ripple marks. At its high stage the lake is roughly estimated to have held more than 500 cubic miles of water of which nearly three-fourths was stored above a constricted part of the Clark Fork Valley called the Eddy Narrows. Calculations based on available incomplete data indicate a flow through the Eddy Narrows that reached a maximum of 9.46 cubic miles per hour. Whether the lake was completely drained at that time has not been determined, but a later set of beaches testifies that the basin held a lake soon after the rapid outflow. Apparently the final draining was gradual.

INTRODUCTION

In the Pleistocene much of the Clark Fork drainage basin in northwestern Montana (Fig. 1) was briefly submerged by "Glacial Lake Missoula" (Pardee, 1910). A lobe of the Cordilleran glacier advanced southward along the Purcell Trench into northern Idaho where it occupied the basin of Pend Oreille Lake and crowded up the valley of the Clark Fork River to a position near the Montana State line. The water thus impounded in the Clark Fork Valley rose to a maximum altitude of 4150 feet or about 2000 feet higher than the floor of the valley just above the dam. On the east, south, and west it was held in by mountains, and on the north possible outflow channels were blocked by other lobes of Cordilleran ice.

The ice dam is thought to have failed, permitting a sudden large outflow. The consequent depression of the water surface in the arm of the lake above the dam caused huge and rapid currents in the narrow valleys and passes connecting this arm with the wider depressions in the lake basin farther east. The records of these currents are described. The interpretation of certain gravel deposits and denuded surfaces in the Perma and Paradise narrows as the effects of such currents differs from that of Davis (1921) who regarded them, respectively, as "moraines" and "ice scoured slopes." Field work by W. C. Alden and the writer has failed to uncover any evidence of glacial origin.

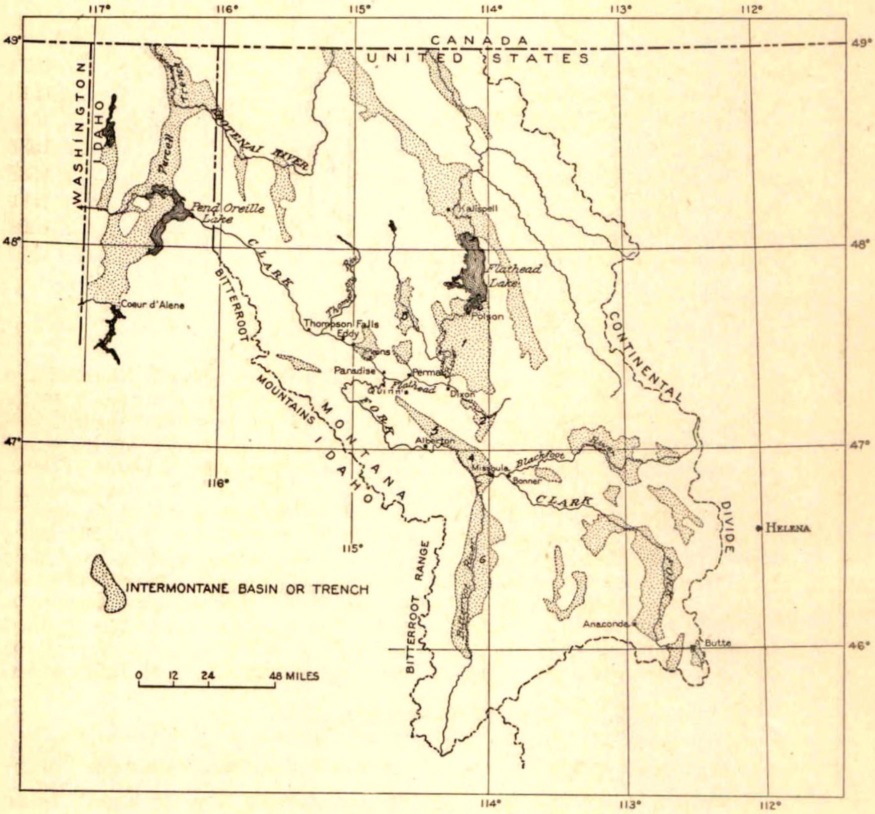

FIGURE 1.—*Index Map*

1. Mission Valley.	3. Little Bitterroot Valley.	5. Ninemile Valley.
2. Jocko Valley.	4. Missoula Valley.	6. Bitterroot Valley.

Not long after the rapid outflowing ceased the basin again held a lake. So far it is not clear whether this body was part of the original ponded water whose escape had been halted or a new filling. In contrast to the earlier outrush, final draining seems to have been very slow.

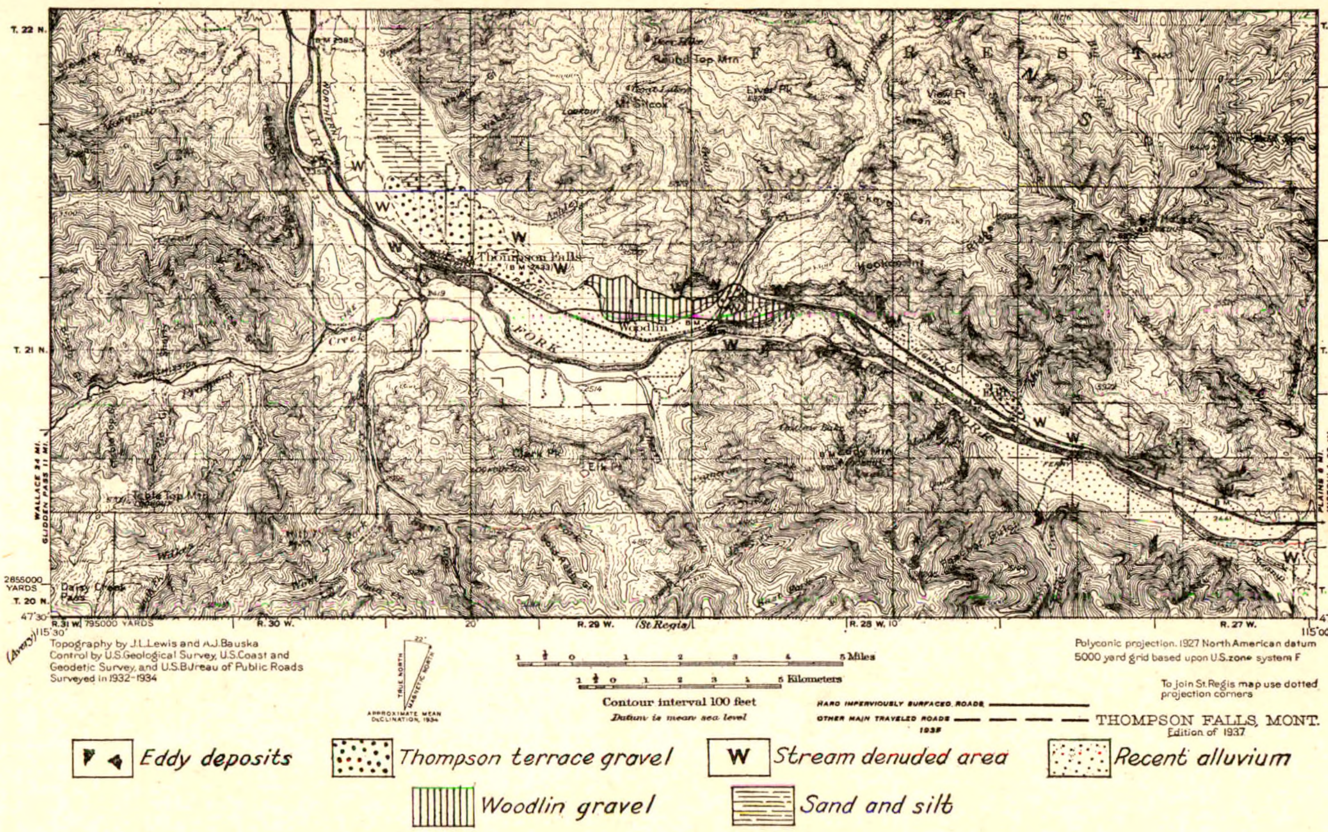

FIGURE 2.—*Eddy Narrows and Thompson Falls Valley*

Reproduced from Thompson Falls, Montana, topographic sheet of the U. S. Geol. Survey. (The term "eddy deposits" in the legend is equivalent to "high eddy deposits" in the text.)

ACKNOWLEDGMENTS

The field work for this paper was facilitated by the use of aerial photographs of the United States Forest Service without which much of the evidence presented could not have been obtained. To W. C. Alden of the United States Geological Survey the writer is indebted for collaboration in the field work and to W. W. Rubey and W. B. Langbein of the same bureau for calculations of stream flow.

TOPOGRAPHY

BASINS

Although the region of Glacial Lake Missoula is occupied chiefly by mountains, the topography is diversified by a number of wide valleys (Fig. 1) or intermontane basins.

The basins are enclosed by generally rough and steep mountains standing from 2000 to 4000 feet higher. Exceptionally high rugged ranges west of the Bitterroot Valley (Bitterroot Range) and east of Mission Valley (Mission Range) rise abruptly to serrate crests 5000 to 6000 feet above the adjacent lowlands. Some of the basins are isolated except for narrow outlet valleys that join them in a common drainage system. Others are more closely connected in irregular groups and in certain linear series known as trenches. Most of the individual basins are one-fifth to half as wide as they are long. Their floor areas range from about 50 square miles in Camas Prairie to 400 square miles or more in Mission Valley and consist mainly of gently sloping plains that merge with a medial flat or flood plain or end as terraces above it. Several of these basins were submerged by the central and eastern parts of the glacial lake. Among them the Mission, Jocko, and Little Bitterroot valleys (Figs. 1, 3) form a closely spaced group drained by the Flathead River. Another group that drains into the Clark Fork River above the mouth of the Flathead includes the Missoula, Ninemile, and Bitterroot valleys. Two other relatively small basins submerged in the central part of the lake are of particular interest in its history: Plains Valley (Fig. 9; Pl. 1, fig. 1) along the Clark Fork River a few miles below the mouth of the Flathead, and Camas Prairie, east of Plains Valley and a short distance north of Perma.

NARROWS

Clark Fork River and its tributaries drain the basins through comparatively narrow valleys which in certain places become severely constricted and gorgelike narrows. The 70-mile stretch of the Clark Fork Valley between the Missoula-Ninemile-Bitterroot group of basins and the Plains Basin generally is not more than 1 or 1½ miles wide and is enclosed by steep, rugged mountains 2000 to 4000 feet high. Parts narrowed to half a mile or less occur near Alberton (Alberton Narrows), above Quinn's Springs (Quinn's Narrows), and below the mouth of the Flathead River from Paradise to Plains (Paradise Narrows). The Clark Fork Valley throughout 10 miles below Plains Basin is again less than a mile wide (Eddy Narrows) between steep, rugged slopes 4000 feet high (Fig. 2). The valley widens abruptly at the lower end of this stretch and from there for about 70 miles to the northwest to the Idaho State line is generally straight and averages 2 and 4 miles in width. The Cabinet Mountains northeast of this

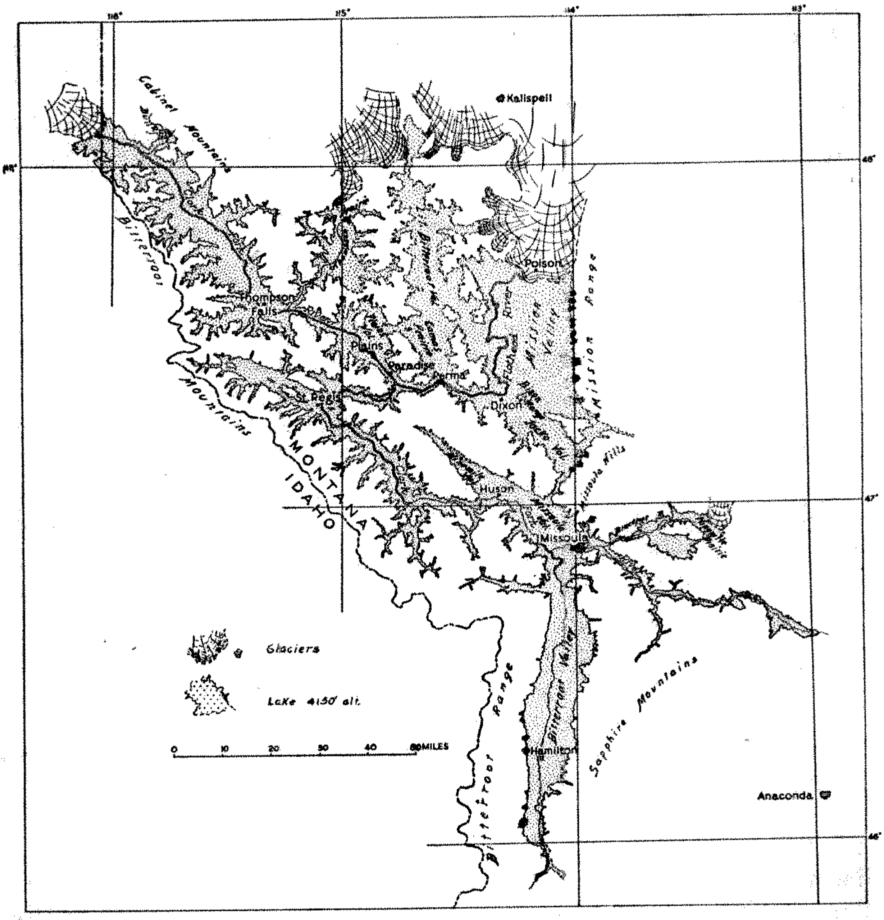

FIGURE 3.—*Glacial Lake Missoula*
High Stage

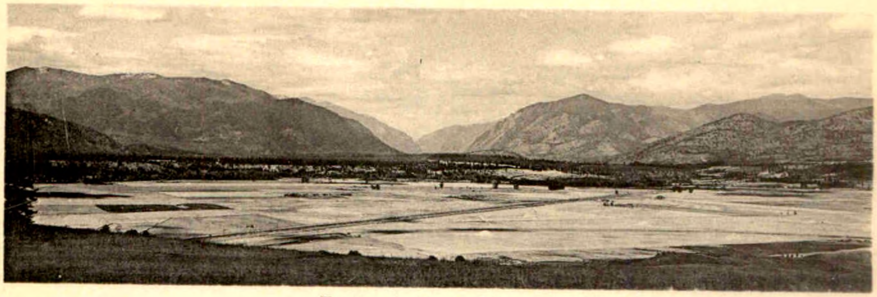

FIGURE 1. PLAINS VALLEY

One of smaller intermontane basins. Looking west into Eddy Narrows, the outlet. Timbered terrace in background is part of coarse gravel deposit. Cultivated area of Recent alluvium.

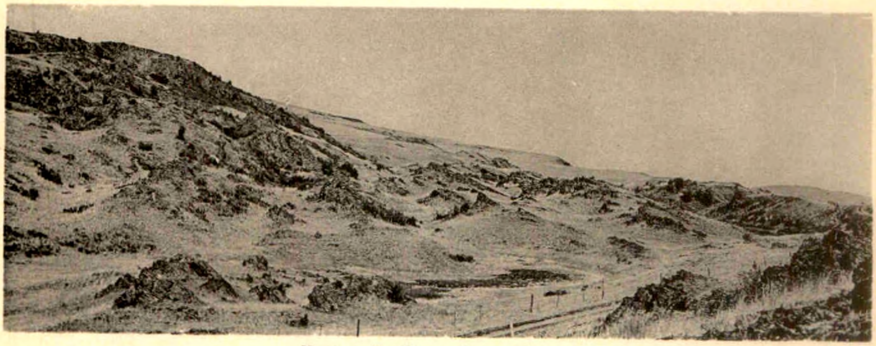

FIGURE 2. WILLS CREEK PASS

Looking north (upstream). Ponded channel depression in foreground. Projecting rock ledges with appended gravel deposits.

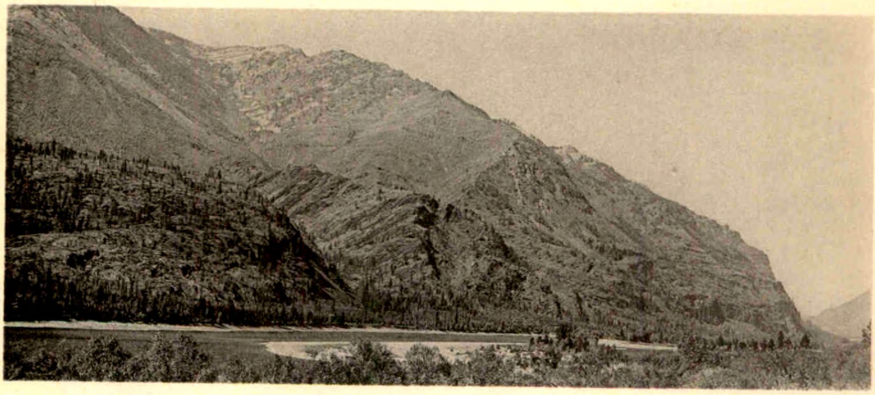

FIGURE 3. SOUTH SIDE OF EDDY NARROWS

Near Outlaw Gulch. Severely denuded spur at left. Normal waste mantle begins above 1000 feet.

FEATURES OF GLACIAL LAKE MISSOULA

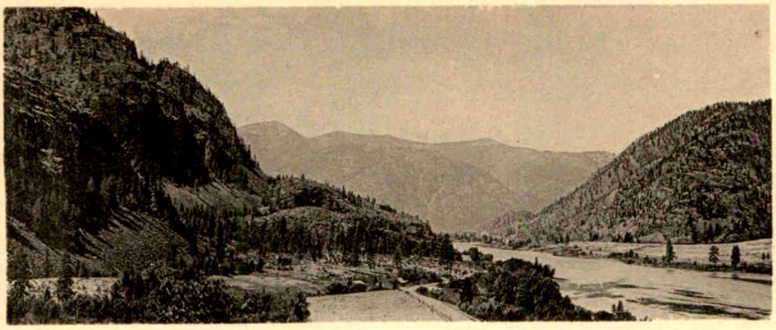

FIGURE 1. LOOKING WESTWARD (downstream)
About 1 mile above Perma. Denuded rock bench at left separated from cliff by trench partly filled with Recent talus.

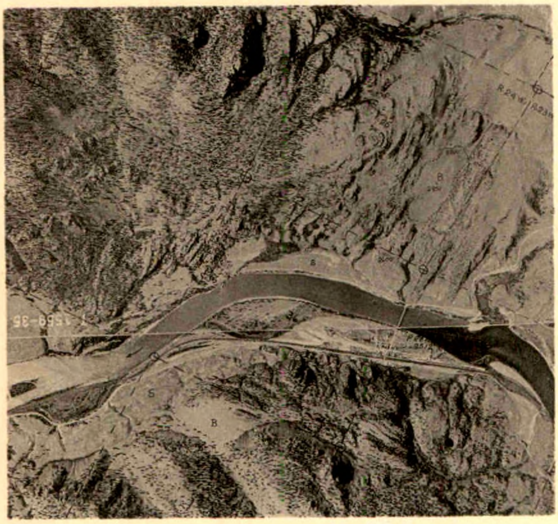

FIGURE 2. ENTRANCE TO NARROWS
Stream-denuded surfaces and channel excavations with ponds (B) and gravel deposits. From aerial photos by U. S. Forest Service.

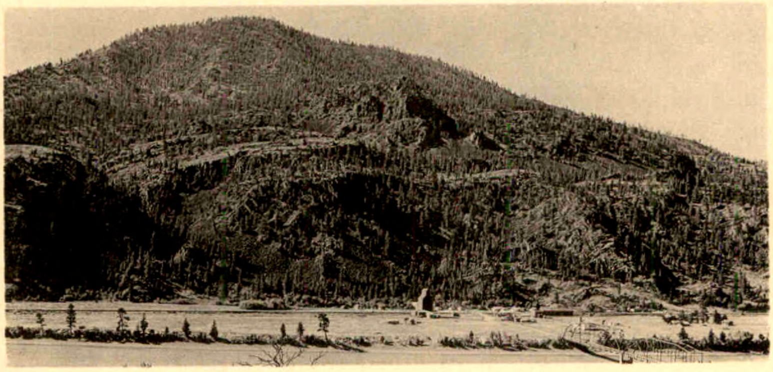

FIGURE 3. DENUDED ROCK BENCH AT PERMA.

PERMA NARROWS AND ROCK BENCH

part of the valley are steep and regular and are 3000 to 5000 feet higher than the valley. The Bitterroot Mountains opposite are lower and less regular. Beyond the State line the Clark Fork Valley opens into the Purcell Trench which in turn joins the Spokane Valley.

The outlet of the Mission group of basins, the valley of the Flathead River, narrows from 2 miles near Dixon to less than a mile at McDonald Station 7 miles downstream (McDonald Narrows). At Perma, 6 miles farther down, the valley becomes still narrower. From Perma to the mouth of the Flathead River, where it merges with the Paradise Narrows, it averages less than half a mile in width (Perma Narrows). As the width of the valley decreases, the enclosing slopes become higher and more rugged. Below Perma they rise 3000 feet or more with slopes characterized by cliffs and bare rock ledges (Pl. 2, fig. 1).

The river bed generally occupies half or more of the valley floor within the different narrows. The remainder of the floor is formed chiefly of low alluvial terraces and an occasional projecting rock knob or ridge. In the adjoining wider stretches of the Clark Fork Valley the terraces are larger and commonly stand 100 to 200 feet above the river. Other pertinent topographic features described in the section on evidences of the unusual currents are wind gaps in the rim of Camas Prairie Basin.

GEOLOGY

BEDROCK FORMATIONS

The region of Glacial Lake Missoula is largely underlain by pre-Cambrian Belt sedimentary formations which, with some intrusive granitic bodies, compose the mountains and extend beneath the Tertiary and later sediments in the lowlands. In the Camas Prairie and Plains basins and the different narrows, the older rocks consist mainly of dark-gray argillites and sandstones of the Prichard formation and Ravalli group, and of a number of diorite dikes and sills. The rocks are deformed by large north to northwestward trending folds and faults and are closely jointed as a rule. Most of the finer-grained beds separate into thin slabs along cleavage planes generally parallel to the bedding. In addition the rocks are traversed by shear zones represented at the surface by notches and trenches, conspicuous on the sides of Perma and Paradise narrows.

These rocks weather to a rubble and finally to a fine-grained sand-clay soil. Above or beyond the area flooded by the lake rubble and soil mantle most of the level and moderate slopes to a depth of several feet, with only particularly resistant beds or dikes projecting through it. Close spacing of joint and cleavage planes produces a rather fine shingle talus with comparatively few large slabs or blocks.

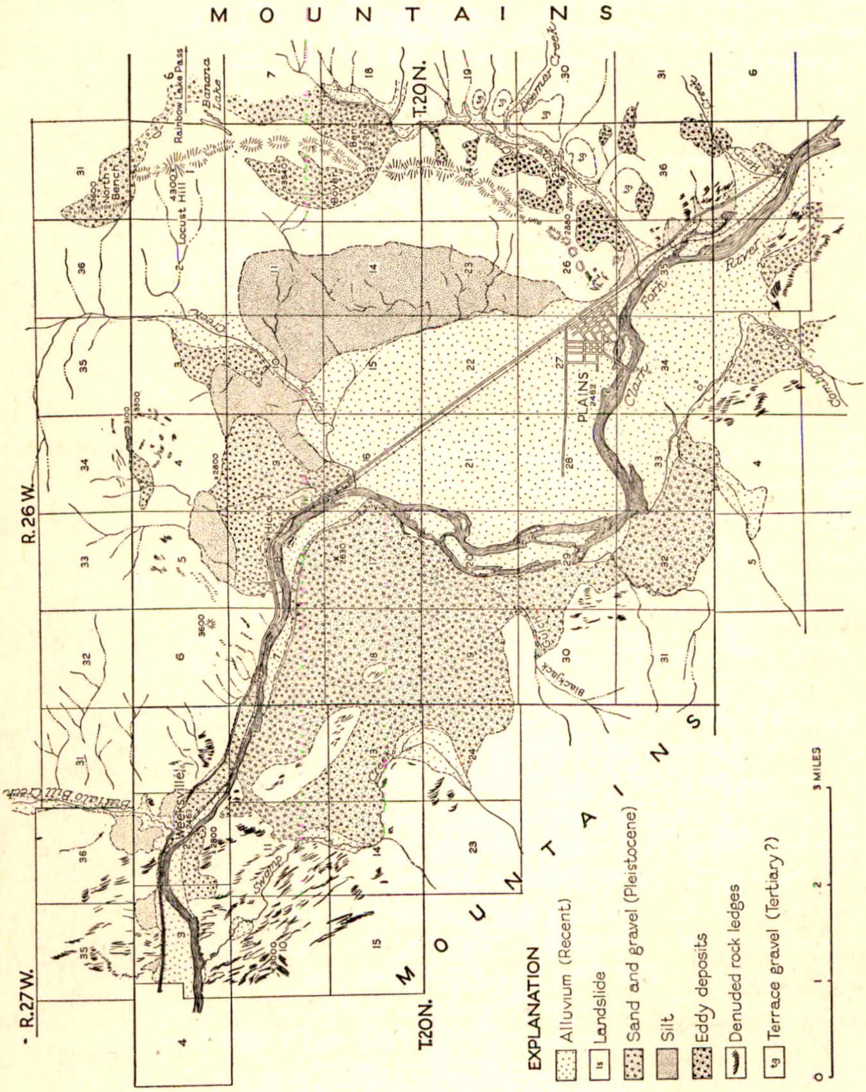

FIGURE 4.—*Southern Part of Plains Basin*

Denuded rock ledges indicated by lines parallel to their trends. The landslide indicated in the legend is on the bank of the river in Sec. 8, T. 20 N., R. 26 W.

On slopes that faced the wider reaches of water the mantle was more or less modified by wave action as described under the section on Beaches. In the narrows the mantle was largely removed by currents during the outflow.

TERTIARY (?) DEPOSITS

Tertiary sediments, extensive in the Missoula-Ninemile and Bitterroot basins, have not been positively identified within the submerged basins to the north and west. In those areas deposits unconformably on the bedrock which may be Tertiary include: (1) water-laid volcanic tuff at the north of Little Bitterroot Basin; (2) "rusty" gravel and sandstone exposed in road cuts and ditches along the foot of the Bison Range; (3) partly decomposed gravel exposed by a road cut in a bench east of Boyer Creek in Plains Basin; and (4) a small patch of gravel with smooth quartzite cobbles on the ridge west of Camas Creek about 2 miles north of Perma.

DEPOSITS ASSOCIATED WITH THE GLACIAL LAKE

Moraines and erratics.—The Pleistocene formations are comprehensively described by W. C. Alden (manuscript in preparation). Among those most closely related to the glacial lake are moraines of contemporaneous mountain glaciers and Cordilleran ice lobes. Several small moraines within the margin of the lake (Fig. 3) were deposited in the Mission and Jocko valleys by glaciers from the Mission Range and Missoula Hills on the east, and in the Bitterroot Valley by glaciers from the Bitterroot Range on the west. Larger moraines were formed at the fronts of the Cordilleran ice lobes that held back the lake on the north. The largest, the Polson moraine at the foot of Flathead Lake, marks the southern limit of the ice which held the arm of the glacial lake in the Mission Valley. Other moraines were deposited by Cordilleran ice lobes in Ninemile Prairie on the Blackfoot River, north of Big Prairie on the Thompson River, and in the valley of Bull River a few miles above its junction with the Clark Fork. Somewhat modified morainal material deposited by the lobe that blocked the Clark Fork Valley is exposed locally near the Idaho State line. Presumably it underlies and thus forms a peculiar sand-covered pitted terrace south of the river at Cabinet Gorge.

Erratics, evidently transported by icebergs from the different glaciers that entered or touched the lake, were distributed irregularly through the submerged areas.

A large body of fine-textured well-washed gravel (Fig. 2) apparently outwash from the retreating ice lobe in Thompson River Valley, forms 50-foot terraces in Clark Fork Valley above and below the entrance of Thompson River.

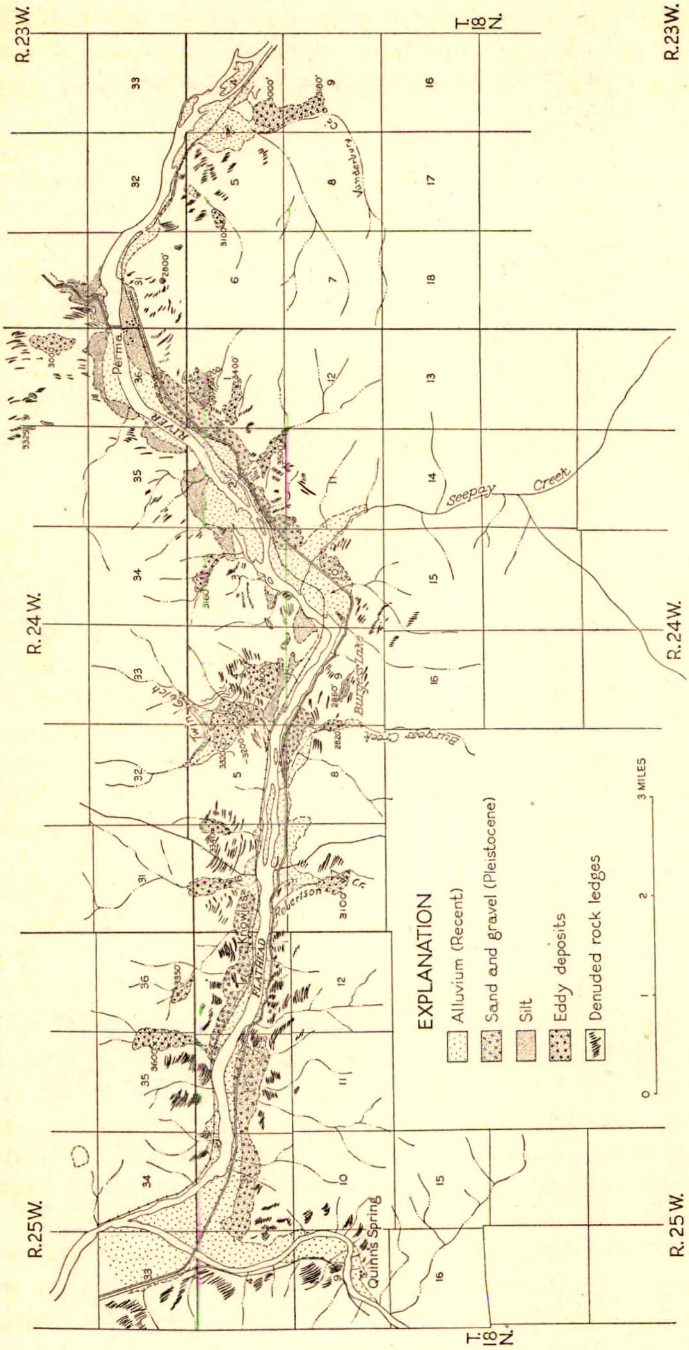

FIGURE 5.—*Perma Narrows*

Denuded rock ledges indicated by lines parallel to their trends.

Silt.—Light-colored varved silt, composed of glacial rock flour derived mostly from the lobes of Cordilleran ice, forms discontinuous terraces 50 to 100 feet high in most of the submerged basins. The silt is confined to the lower parts of the basins and thins out upslope at the height of the terraces which evidently mark its original upper limit. The Bitterroot Basin and, above the mouth of the Blackfoot River, the valley of the Clark Fork received meltwater from local mountain glaciers only, and in both areas little silt was deposited. In the Missoula-Ninemile Basin, however, silt derived apparently from large glaciers in the Blackfoot Valley forms extensive terraces 50 feet or more high between Missoula and Huson. In the Jocko Basin comparatively small discontinuous

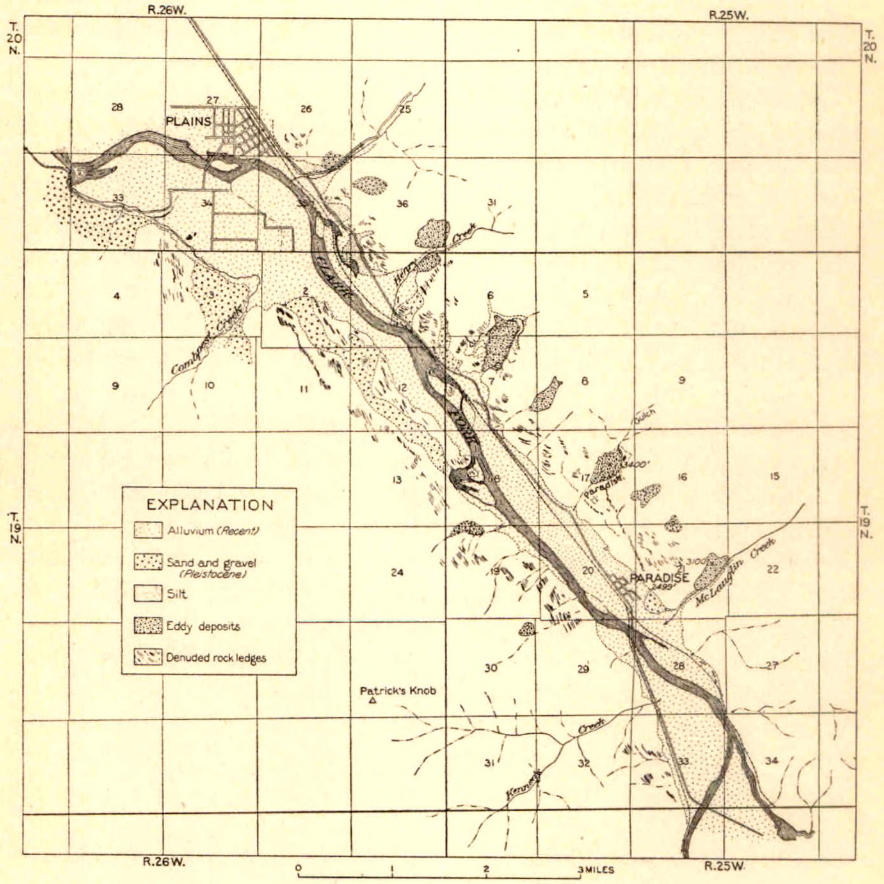

FIGURE 6.—*Paradise Narrows*

Denuded rock ledges indicated by lines parallel to their trends.

terraces are composed of silt that came chiefly from glaciers washed by the arm of the lake in upper Jocko Valley. Silt discharged by the lobe of Cordilleran ice that filled the basin of Flathead Lake occupies large areas in the lower parts of the Mission and Little Bitterroot basins and forms discontinuous terraces up to 50 feet in height in the valley of the Flathead River (Pl. 7, fig. 3) to the Perma Narrows.

The silt in Little Bitterroot Basin is exposed by stream trenching to a depth of about 100 feet and is found to 200 feet in the middle of the basin in wells (Meinzer, 1916). In the Plains Basin, silt carried from either or both the Mission and the Missoula-Ninemile basins forms terraces about 100 feet high in a large area north of the river (Fig. 4).

In the Perma and Paradise narrows limited exposures of the silt occur (Figs. 5, 6) in protected places, and in the other narrows none has been observed. The somewhat wider stretches of the Clark Fork Valley between the Alberton Narrows and Quinn's Narrows and from the Eddy Narrows to the Idaho State line appear to contain silt in places, but its extent and relations have not been determined.

Deposition of the silt apparently was limited to the lower parts of the basins by conditions similar to those observed at Lake Mead above Boulder Dam on the Colorado River and in other reservoirs; owing to its higher specific gravity, slowly moving turbid water sinks and flows along the bottom without mixing with the overlying clarified water (Grover, *et al.*, 1938). In Lake Missoula the silt, in order to reach Plains Basin, must have been thus carried 50 to 75 miles from its sources in glaciers discharging their meltwater in the Mission and Missoula-Ninemile basins. No silt was deposited in Camas Prairie because none of the entrances to that basin were low enough to divert the layers of turbid water passing down the adjacent Little Bitterroot Basin and the Perma Narrows.

In Little Bitterroot Basin upper and lower silt formations separated by sand layers, and, at one place by gravel, are indicated by well records (Meinzer, 1916, p. 15). The partings occur at depths of 80 to 150 feet, and the underlying silt is 100 feet or more thick.

Deltas.—At all the different still stands of the lake, deltalike deposits of alluvium were doubtless formed at the upper limits of the arms extending into the many tributary valleys. The level of the water surface shifted rapidly so that individual deposits generally must have been rather small. Also, as they were exposed to later stream attack, few would remain sufficiently unmodified to be recognized. Gravel deposits in the Vermilion River Valley that may be the remnants of exceptionally large deltas begin at Cataract Creek about 5 miles from the mouth of the valley and at Sims Creek about 3 miles farther up. The deposit at Cataract Creek forms a terrace 500 feet high along the south side of Vermilion River. From an altitude of 3200 feet at the lower end the surface ascends gradually upstream. The deposit consists of gravel overlying glacial till, which was observed at one place by W. C. Alden (personal communication). On the opposite side of Vermilion River and about a mile above Cataract Creek is a similar terrace in Lyons Gulch. A small undrained area at the back of this terrace indicates that the deposit was formed by currents turning out of the valley of Vermilion River. The terrace at Sims Creek is about 300 feet high, and the surface rises gradually upstream from an altitude of about 3550 feet at the terrace front.

A gravel filling in McLaughlin Gulch east of Paradise apparently was formed when the lake surface was at 3200 feet. About 1½ miles from the mouth of the gulch this deposit begins as a terrace about 30 feet high. Upstream it narrows and ends within half a mile because of the more steeply ascending bedrock channel of the gulch. The deposit is partly trenched by the stream and shows rather vague top-set bedding. Remnants of other gravel fillings apparently related to still stands of the lake extend up the valley of the St. Regis River from points above Taft at 3800 feet altitude and near Borax at 4100 feet. Both deposits rise gradually upstream from low terrace fronts.

Beaches.—At many places on the untimbered sides of the basins up to an altitude of 4150 feet the horizontal markings of closely spaced lake beaches are plainly visible from moderate distances. Within 1000 feet of the limiting altitude as many as 40 may be counted in favorable exposures (Pl. 5, fig. 1). Thus different still stands were so closely spaced that practically all of a given slope below 4150 feet was exposed to the waves which were most effective, of course, on the higher parts of a slope because they faced the wider stretches of water.

Scattered observations indicate that the beaches consist of a generally slight terracelike indentation thinly covered by cleanly washed and sorted gravel. On slopes that faced the wider arms of the lake the terrace or bench, as it rounds projecting spurs, may be several yards wide, and its related bars in adjacent re-entrants may be measured in cubic yards. Other conditions being equal, the beaches are not only more extensive around the larger basins but appear to be more deeply indented around their southern and eastern sides as if the prevailing winds were from the north and west. In many places rows of loose stones are ridged along the backs of the beaches where, presumably, they were shoved by expanding lake ice.

The beaches are less prominent in the smaller basins and become inconspicuous or absent in the valleys that held the narrower arms of the lake. In the relatively narrow valley of the Clark Fork River above Missoula they may be detected on hills west of Harvey Creek and west of Tyler Creek at altitudes between 3900 and 4150 feet. In the somewhat wider valley below Thompson Falls beaches may be observed between 2700 feet and 3500 feet altitude on partly bare slopes south of Tuscor and northeast of Noxon. Within the narrows beaches have not been identified.

RECENT ALLUVIUM

Within the basins recent alluvium forms flood plains and low terraces along the streams. In the narrows and the relatively narrow stretches of Clark Fork Valley, particularly below Thompson Falls, such features are absent or negligible because the river generally occupies a steep-sided inner trench.

EVIDENCES OF THE UNUSUAL CURRENTS

CHANNELED WIND GAPS IN RIM OF CAMAS PRAIRIE BASIN

General statement.—Camas Prairie Basin north of Perma is almost 12 miles long from north to south and 5 or 6 miles wide in the middle. Its floor descends gradually from an altitude of about 3100 feet at the north to about 2900 feet at the south and includes a flat from which the surface rises gradually to the moderately steep slopes of the surrounding mountains. At the high stage of the glacial lake (Fig. 3), the mountains on the east and west sides of the basins stood well above the water, but those on the north and south were almost entirely submerged. The arm of the lake in this basin was confluent, therefore, with the arms in Little Bitterroot Basin on the north and Perma Narrows on the south. In addition a strait in Rainbow Lake Pass connected it with the arm in Plains Basin on the west. When the water surface fell 500 feet or to about 3650 feet this pass became dry, and the northern rim partly emerged, thus dividing the water connection to the Little Bitterroot arm among four gaps. As the lake shrank further these gaps became dry, Duck Pond and Big Creek passes at about 3500 feet altitude, Markle Pass at 3400 feet, and finally Wills Creek Pass at 3300 feet. A wide gap at the south emerged when the lake fell about 300 feet more, and, except for the narrow outlet trench east of this gap, the water connection between the basin and Perma Narrows ceased.

Gaps in north rim.—Although they begin and end as it were "up in the air", the four gaps in the northern rim of Camas Prairie Basin carried large and vigorous streams. In Duck Pond Pass a marshy pond with coarse gravel beyond it, and a bare rock hump, resemble a basin and succeeding bar formed by a south-flowing current plunging over the obstructing hump. Markle Pass and Wills Creek Pass have channel excavations and discontinuous gravel deposits associated with projecting knobs and ridges. In Markle Pass (Fig. 7) an undrained depression 50 feet deep adjoins a prominent knob to which a gravel train is attached on the downstream side. East of the knob Recent talus partly fills another channel, and beyond that, to a height of 100 feet or more, the surface is denuded and channeled. South of the pass

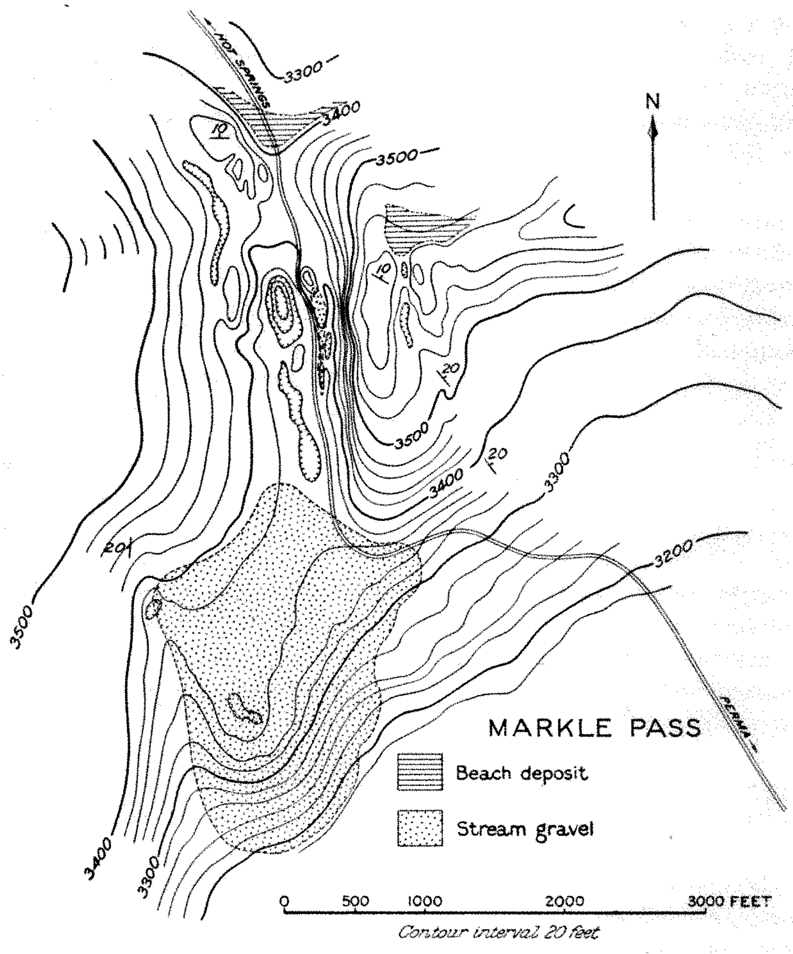

Figure 7.—*Topography and geology of Markle Pass*
Areas not patterned represent tilted Belt rocks.

coarse gravel with scattered large angular fragments plucked from ledges within has the form of a bar about 50 acres in extent built out to a height of 150 feet over the slope descending to the basin.

In Wills Creek Pass, channel features are developed on a still larger scale. Along the base of a steep slope at the east side is a pond 1000 feet long. On the west side (Pl. 1, fig. 2) are channels with undrained hollows and bare angular knobs and ridges projecting above discontinuous deposits of gravel. South of the pass a deltaic bar is built out for 1 mile on the slope descending to the basin (Pl. 3). The bar is 200 feet high at the southern end and like the deposit in front of Markle Pass is composed of coarse gravel with large angular fragments plucked from ledges within the channel. In front of Big Creek Pass a relatively small, almost circular barlike gravel deposit is built over a steep slope.

Apparently the inflowing currents were caused by a lowering of the water surface in the arm held by Camas Prairie Basin relative to the surface in Little Bitterroot Basin. Since the channels show no after effects of dwindling currents the flows must have ceased rather abruptly. The evidences described indicate actively eroding currents at least 30 or 40 feet deep in Duck Pond and Big Creek passes, 150 feet deep in Markle Pass, and 250 feet deep in Wills Creek Pass.

Gap at south.—The south rim of Camas Prairie Basin is lowered by a wide gap to within a few hundred feet of the adjacent basin floor and is penetrated east of the gap by a narrow steep-sided outlet trench. Evidences within the gap of a huge outflowing stream include denuded rock surfaces, discontinuous gravel deposits, and basins, some holding bonds or marshes; all are characteristic of stream beds in rocky channels (Pl. 2, fig. 2). One of the larger gravel deposits (B on Pl. 2, fig. 2) rises from a hollow below an obstructing rock hump to a rounded bar. It extends south to the top of the slope descending to the river opposite Perma where it takes the form of a high eddy deposit, a result, apparently, of the commingling of the stream from the gap with that flowing down the Perma narrows.

Rainbow Lake Pass.—Particularly impressive evidences of stream action appear in Rainbow Lake Pass, a steep-sided trenchlike valley (Pl. 4) which extends, at an elevated position, from the northwest corner of Camas Prairie Basin through the dividing range to Plains Basin on the west.

General Form and Relations: The Camas Prairie side of the pass has two entrances that converge near the mountain axis, about a third of the way across. The northern entrance, which is traversed by the

Plains to Elmo automobile highway, is about half a mile wide, and the southern entrance is somewhat narrower. Both commence at an altitude of about 3650 feet at the heads of small valleys that ascend gradually from the floor of the basin some 700 feet below. The pass is about 6 miles long. Its profile descends westward a total of about 50 feet most of which takes place in the branches just above the junction.

Rainbow Lake (Pl. 6, fig. 1) occupies a depression in a narrowed stretch of the pass just below the junction of the two branches. Allowing for small seasonal fluctuations its surface averages about 3600 feet in altitude as determined by aneroid. The lake is fed by two small streams, has no visible outlet, and the water remains fresh.

The pass widens from about half a mile at the lake to nearly a mile where it comes out of the mountains to Plains basin at the west. A view of it from points within the basin, however, is obstructed by a prominent elevation called Locust Hill standing directly in front of the entrance. From Locust Hill a low foothill ridge, parallel to the mountains east of the basin, extends south to the Clark Fork River at Plains (Fig. 4). Between this ridge and the mountains is the relatively narrow valley of Boyer Creek.

South of Locust Hill shelflike terraces are perched along both sides of the foothill ridge within 100 feet of the top. They are about 3550 feet in altitude, 1½ miles long, and toward the south, where the dividing ridge becomes lower, they approach a gap and meet to form a narrow junction. The terrace (Pl. 4) on the western side of the ridge overlooks Plains Basin from a height of 1000 feet. At its southern end the terrace on the eastern side of the ridge is built out about half way across the valley of Boyer Creek and stands about 500 feet above the stream bed. Northward it widens until it extends across the valley. Together these terraces compose Boyer Bench. The Plains to Elmo automobile highway goes up Boyer Creek and gradually ascends the eastern front of the terrace.

Channel Erosion: Rainbow Lake Pass exhibits strikingly the effects of a former large and rapid westward-flowing current. Above the lake the floors of both branches have been severely scoured, exposing the bedrock which remains bare except for a thin rubble of its own fragments loosened by Recent weathering (Pl. 6, fig. 2). The floor is furrowed with channel depressions, several of them holding ponds or marshes (Pl. 4). Upstream angles of prominent knobs are slightly rounded as by water. To a height of 200 feet or more prominent surfaces on the side slopes are bare, though farther up similar surfaces are smoothly covered. The depression occupied by Rainbow Lake resembles a hollow or "plunge pool" that would be excavated below a rapids by the combined currents from the two branches.

GRAVEL DEPOSIT: Below the lake and along its south side a deposit of coarse stony gravel covers the floor of the pass. Beyond the pass this deposit fills the space to Locust Hill, where it divides into a smaller branch—extending through a gap at the north and expanding in a terrace at the edge of Plains Basin—and a larger branch that turns south into Boyer Creek Valley to form the eastward-facing terrace of Boyer Bench. The entire deposit is poorly sorted, slightly to moderately water-worn pebbles, cobbles, and boulders, most of them between 2 inches and 1 foot in diameter with a variable but generally scanty proportion of smaller pebbles, sand, and silt but no clay. The deposit is not cemented and but slightly compacted. Slightly water-worn to angular fragments, some of them 5 or 6 feet in length, are widely and irregularly scattered through it. A layer 2 or 3 feet thick of "dry gravel," an aggregate of loose stones without sand or other fine matrix, is spread over much of the surface. Included in this layer are fragments of rounded cobbles presumably split apart by frost action since they came to place.

Recent highway excavations along the east front of Boyer Bench expose below the "dry gravel" or other superficial layers a more or less distinct downslope or fore-set bedding (Pl. 8, fig. 1). In some places the beds are lenslike and composed of relatively small pebbles and cobbles in an abundant sandy matrix. In other places the beds are irregular, relatively coarse, and almost without finer material, thus largely consisting of voids. Except for an occasional erratic, the gravel consists of rocks like the outcrops in Rainbow Lake Pass. Most of them are sandstone and argillite of the Prichard formation and Ravalli group which, because of a slaty parting, tend to separate into slabs. A few are blocky fragments of diorite.

Throughout its area the gravel deposit exhibits the surface characteristics of alluvial stream beds. Of these the minor features include ridges, channels, and basins mostly 5 to 10 feet high or deep but which locally reach 40 feet. The larger features are barlike forms. One of these rises gradually from the lower end of the lake to a height of 70 or 80 feet (Pl. 4). The bar is banked along the north side of the pass opposite a denuded slope on the south side that rises from a channel depression, partly filled with Recent alluvium. Another bar is heaped against the south spur end at the entrance to the pass, a position opposite the scoured east side of Locust Hill.

The ridge at the south of Locust Hill gives evidence of having been overflowed by a large stream from the east. Material eroded from it forms the western part of Boyer Bench. The surface of this deposit is somewhat roughened with small channels and basins except along the front where it is smoothed by a deposit of later beach gravel.

These features apparently were formed by a huge current swinging from side to side as it moved swiftly through the pass eroding the outsides of bends and depositing gravel bars on the insides. Beyond the pass the stream severely scoured the opposing slope of Locust Hill. There it divided and a smaller part passing around to the north formed an embankment at the edge of Plains Basin. The larger part turned south into the valley of Boyer Creek, deposited the east-facing terrace of Boyer Bench, and, at the same time, overflowed the dividing ridge and formed the terrace facing Plains Basin. The flow was of brief duration, and both branches followed the tendency of a stream passing into deep water to build out a deltaic bar as the channel bed falls away. Presumably the flow started when the surface of the arm of the glacial lake in Plains Basin fell below the surface of the arm in Camas Prairie Basin. It ceased when equilibrium was restored between the two basins or when the water surface in both fell below the pass.

Giant ripple marks of Camas Prairie.—Camas Prairie is distinguished by a number of unique ridges that have the form, structure, and arrangement of ordinary ripple marks but are so large that the term "ripple mark" seems inappropriate. It is retained, however, for want of a better one. Except for a few inconspicuous ones below Big Creek Pass, the ridges (Pl. 3) are grouped in belts appended to the other three northern gaps—namely, Duck Pond, Markle, and Wills Creek passes, a relationship suggesting formation by currents entering the basin through those channels. The relative scarcity of similar ridges below Big Creek Pass might be explained by the relatively small size of its stream.

The distribution, linear form, and arrangement of the ripple-mark ridges as mapped were obtained from aerial photographs (by the U. S. Forest Service) of the belt below Duck Pond Pass, and by pace and compass traverses of the belts below Markle and Wills Creek passes. An unmapped part of the Wills Creek Pass belt extends 1 mile or more farther south. The ridges range from 1 foot or less to 50 feet in amplitude and from a few feet to as much as 500 feet in wave length, but the mean height is between 15 and 30 feet and the wave length about 250 feet. Most of the smaller ridges occur toward the lower (basinward) ends of the belts, and the exceptionally high ones near the middle. Toward the lower end of the belt in front of Duck Pond Pass the ridges gradually decrease in size until they fade out completely.

Individual ridges are from a few rods to half a mile or more in length. Their axes are slightly curved, and convex forward (toward the basin). Their crests are rather wide and rounded and with few exceptions their cross profiles are asymmetrical, the southern or downstream sides being the steeper. As exposed in a pit in one of the ridges about 3½ miles south of Wills Creek Pass (near the southeast corner of sec. 12, T. 19 N., R. 24 W.) the gravel is largely subrounded pebbles ranging from 8 to 20 millimeters in diameter. Sandy matrix is rather scant, the mass is loosely compacted, distinctly stratified, and throughout the cross section the dip of the bedding is parallel to the steeper south (downstream) side (Pl. 3). Toward the pass the gravel of the ridges coarsens, but no other exposures of the structure were seen. In comparable situations the ridges in front of Markle and Duck Pond passes are of similar composition and texture.

The giant ripple marks lie on a smooth bedrock surface that rises gently from the interior flat to the bordering mountains, and, according to the direction which they happen to take across this surface, their profiles are sloping or horizontal. Toward the passes they are overlain by the bar-like deposit of coarse gravel at the end of each channel. All appear to have been formed by large currents plunging over the north rim and continuing with decreasing velocity out into the basin. Such currents might have formed when the water surface in Camas Prairie Basin first became lower than the water surface in Little Bitterroot Basin thus causing masses of water to pass over the ridge, chiefly through the gaps. The initial direction and momentum of the descending masses of water plus increased specific gravity due to the fine sediment picked up on the way would tend to hold the currents to the bottom for a considerable distance.

Giant ripple marks in plains basin.—Gravel ridges that resemble the giant ripple marks of Camas Prairie and are probably similar in origin appear in two places at the eastern side of Plains Basin below Rainbow Lake Pass (Pl. 4, Sec. 12, T. 20 N., R. 26 W., Sec. 36, T. 21 N., R. 26 W.). They are relatively inconspicuous and were first observed in the aerial protographs. The larger ones (Sec. 12, T. 20 N., R. 26 W.) form a group in the path of the stream which overflowed the ridge south of Locust Hill. They are 10 to 20 feet high, 200 to 500 feet apart, and their axes are normal to the course of a current descending to Plains Basin. A group of lower and less definite ridges (Sec. 36 T. 21 N., R. 26 W.) bear a similar relation to the stream that reached Plains Basin north of Locust Hill.

Stream-denuded surfaces in narrows.—Mention has been made of rock surfaces in the gaps around Camas Prairie Basin, in particular Rainbow Lake Pass, that appear as if washed bare of waste mantle except for

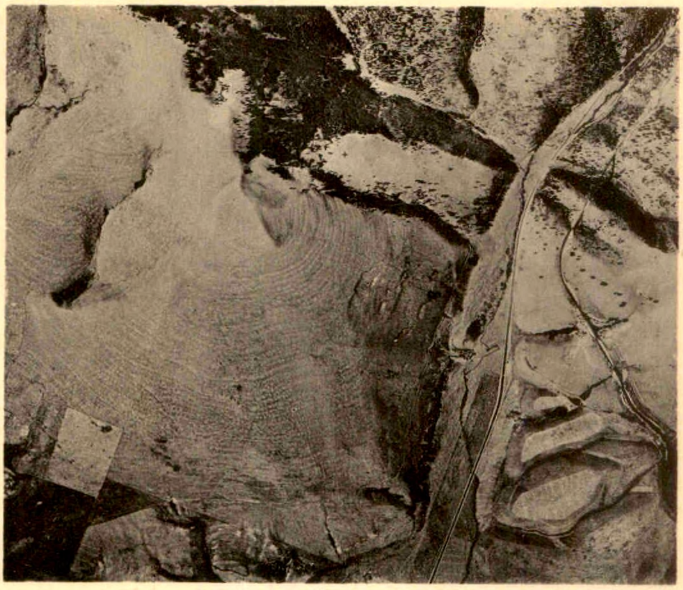

FIGURE 1. BEACHES OF GLACIAL LAKE MISSOULA
Northeast side of Missoula Valley at O'Keefe Creek. Hill is about 1000 feet high, summit is slightly above 4150-foot beach which marks highest lake stage. Aerial photo by U. S. Forest Service.

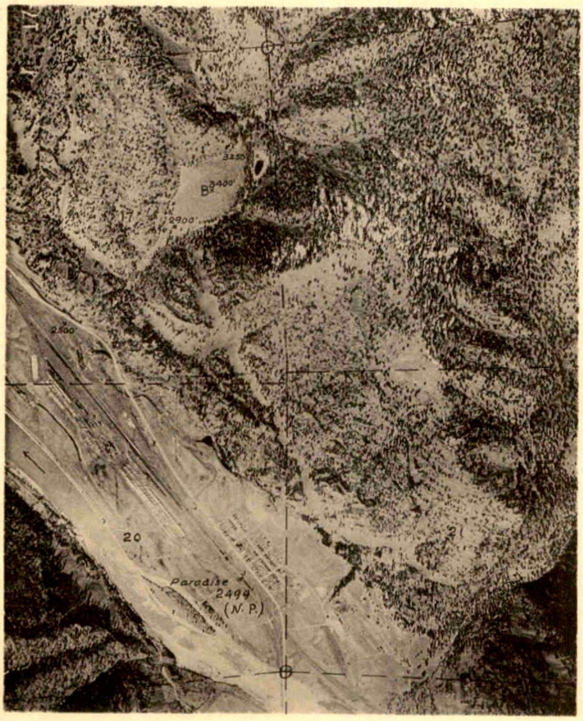

FIGURE 2. PARADISE NARROWS
High eddy deposit with pond at back outlined at B. Light rectilinear streaks are trenches along major joints or shear zones and partly filled with Recent talus. Aerial photo by U. S. Forest Service.

BEACHES AND HIGH EDDY DEPOSIT

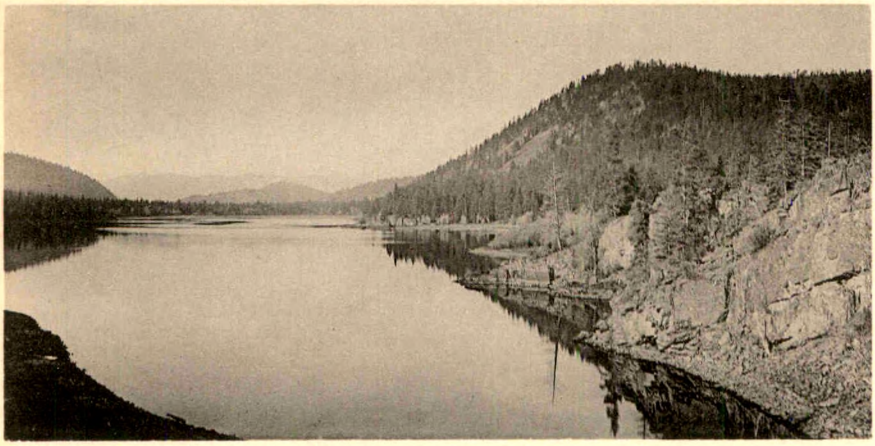

FIGURE 1. RAINBOW LAKE
Fills depression in floor of the pass.

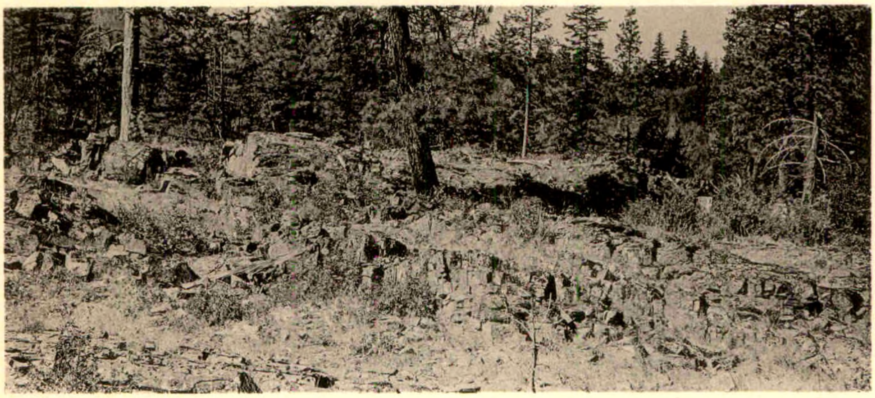

FIGURE 2. BEDROCK FLOOR
Surface denuded of waste. Joints opened by weathering.

RAINBOW LAKE PASS

Recent rubble. A similar condition on a much greater scale appears in the narrows along the Clark Fork and the Flathead rivers through which the ponded waters of the Missoula and the Mission groups of basins escaped. In these sluiceways all humps, knobs, and other convexities of the bedrock that project through the sediments of the valley floor or stand out on the adjoining slopes are manifestly bare of a normal waste mantle to heights varying from several hundred to a thousand feet or more. In contrast, comparable slopes at higher altitudes are smoothly covered, as a rule, to a depth of several feet. Thus in the Alberton Narrows below the Missoula group of basins, prominent surfaces on the side slopes are bare several hundred feet up. Knobs and ridges that project above the present flood plain including rock benches 200 feet high on the south side of the river at the entrance to the narrows, and also just below Alberton, are largely denuded of residual mantle or soil. Farther down, in Quinn's Narrows, similar conditions exist.

To the north in the valley of Flathead River denuded rock surfaces begin to show conspicuously along McDonald Narrows, the first noteworthy constriction of the valley below the Mission group of basins. There a spur projects from the north side of the valley, bare of waste to a height estimated at 1000 feet, the slope above being normally covered (Pl. 7, fig. 3). At Perma, 6 miles farther down, the surface mantle is scoured from a rock bench 400 to 600 feet high (Pl. 2, fig. 3) exposing the small scale irregularities of the rock surface as well as several larger trenches and a basin that holds a pond (Pl. 2, fig. 2).

Continuing down through the Perma and Paradise narrows into Plains Basin and on through the Eddy Narrows (Pl. 1, fig. 3) prominent rock surfaces are likewise denuded to a height of 1000 feet or more above the river.

Along the somewhat wider stretches of Clark Fork Valley between the Alberton Narrows and Quinn's Narrows similar conditions are less noticeable, but even there a normal waste mantle is generally absent from the more prominent knobs and spurs. The moderately wide stretch of valley from Thompson River down to the Idaho State line has not been examined in detail, but the lower slopes of Tuscor Hill and other projecting spurs are noticeably bare.

High eddy deposits.—Sediment from confined turbulent streams active enough to keep their channels clear is commonly deposited in recesses along the sides by eddying currents. The gravel bodies described below are thought to be of such an origin. They are like the eddy deposits of ordinary streams that are confined to steep-sided channels, but their production calls for streams of unusual magnitude.

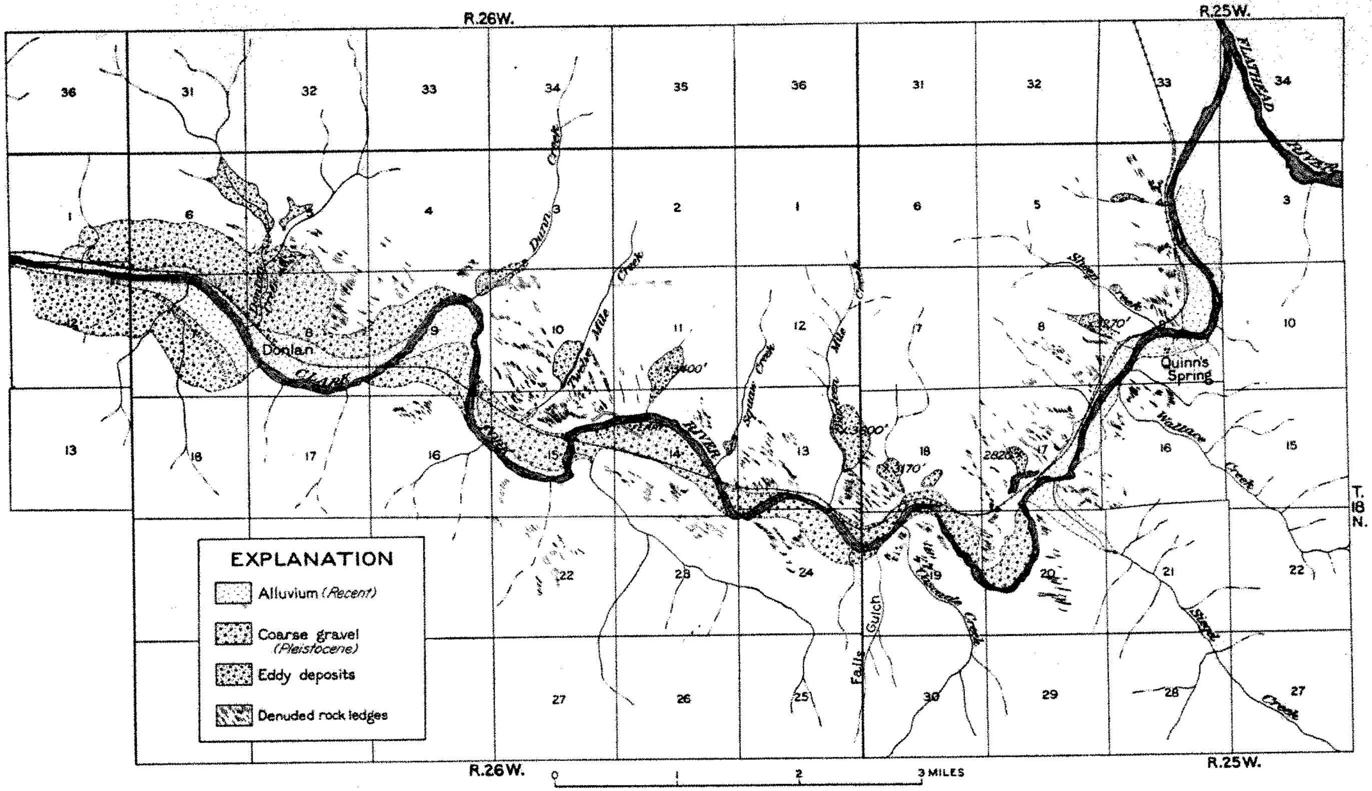

FIGURE 8.—*Quinn's Narrows*

Denuded rock ledges indicated by lines parallel to their trends.

The high eddy deposits occur in side gulches, pockets, or other entrants along the sides of the different narrows. The locations of a number and the heights they attain, as approximately determined by aneroid barometer or triangulation with a Brunton compass, are given in Table 1.

TABLE 1.—*Location and height of High Eddy Deposits in the Clark Fork and Flathead River Valleys*

No.	Location	Height above valley floor (feet)
	PERMA NARROWS (FIG. 5)	
1	Vanderburg Creek (secs. 4, 9)	680
2	E. of Perma (sec. 5)	600
3	W. of Perma (N. ½ sec. 1)	700
4	W. of Perma (center sec. 1)	900
5	W. of Perma (sec. 2)	1000
6	N. of Seepay Creek (sec. 3)	660
7	Burgess Creek (secs. 8, 9)	320
8	E. of Twin Gulch (sec. 4)	300
9	Twin Gulch (Stouts Bar) (secs. 4, 5)	820
10	Robertson Creek (sec. 7)	600
11	E. of Knowles (secs. 5, 6)	840
12	E. of Knowles (sec. 6)	1100
13	NW. of Knowles (sec. 36)	856
14	W. of Knowles (sec. 35)	1200
	PARADISE NARROWS (FIG. 6)	
15	McLaughlin Creek (sec. 21)	610
16	Paradise Gulch (sec. 17)	920
17	W. of Paradise Gulch (sec. 8)	970
18	E. of Henry Creek (secs. 6, 7)	1120
19	Henry Creek (secs. 1, 36)	630
	QUINN'S NARROWS (FIG. 8)	
20	Opposite Quinn's Springs (secs. 8, 9)	790
21	Opposite Siegel Creek (sec. 17)	320

TABLE 1.—*Location and height of High Eddy Deposits in the Clark Fork and Flathead River Valleys—Continued*

No.	Location	Height above valley floor (feet)
22	Opposite Cascade Creek (sec. 18)	575
23	Fourteen Mile Creek (sec. 13)	1035
24	Opposite CCC Ferry (sec. 11)	855
25	Twelve Mile Creek (sec. 10)	645
26	Dunn Creek (secs. 9, 10)	325
27	Donlan Creek (secs. 5, 6)	625
	EDDY NARROWS	
28	S. side below Weeksville	750
29	Poacher Gulch	650
30	Quartz Creek	1000
31	Malone Creek	675
32	Under Kookoosint Ridge	1075

Similar deposits are present in several recesses in Boyer Creek Valley (Fig. 4) along the east side of Plains Basin below Rainbow Lake Pass. More detailed examinations of the different narrows, particularly the more thickly timbered sides, would doubtless reveal others.

The high eddy deposits are characterized by a steep front slope, a rounded top, and a level or gentle back slope as illustrated, for example, by the deposit (No. 16, Table 1) in Paradise Gulch. This body, set somewhat back of the spur end cliffs, begins about 300 feet above the river and rises at a 30-degree slope 600 feet more to a wide rounded top at an altitude of 3400 feet (aneroid). Toward the back its surface descends gradually 100 feet or so to an undrained depression containing a pond (Pl. 5, fig. 2). Back of the depression the buried gulch channel reappears. Another body (No. 18) in the second gulch to the west reaches an extreme height of 1120 feet (about 3500 feet altitude). The profile of its top rises gradually to the west. At the back is a small flat. Still another deposit (No. 9) that is very conspicuous (Pl. 7, fig. 1) from the highway in the valley fills out the lower part of Twin Gulch even with the valley side, its rounded top being about 800 feet above the river. The back slope descends to an alluvial flat drained by a trench at the east.

High eddy deposits formed in gulches with active streams have been more or less washed away. One in the valley of Henry Creek (No. 19) has been separated into two parts, and another (No. 12) in a gulch east

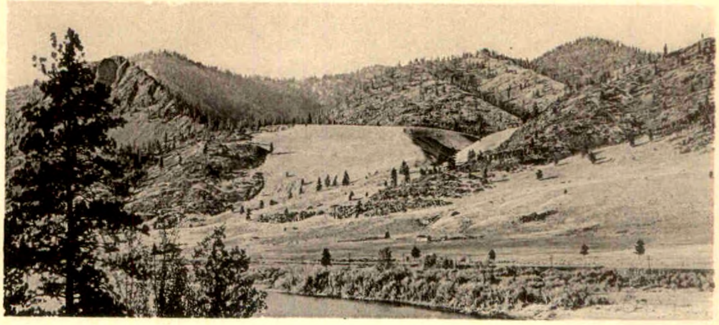

FIGURE 1. STOUT'S BAR
In Twin Gulch on north side of Perma Narrows 2 miles east of Knowles (Fig. 5). Top rises gradually toward left. Maximum height above river about 800 feet. Adjoining rock slopes on right and left cleanly scoured of waste.

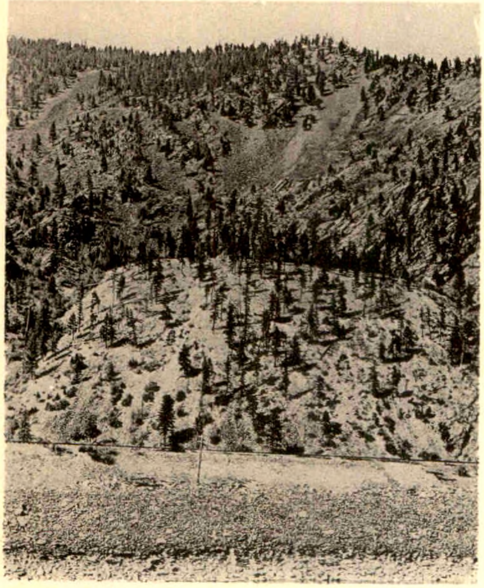

FIGURE 2. NORTH SIDE OF QUINN'S NARROWS OPPOSITE SIEGEL CREEK.

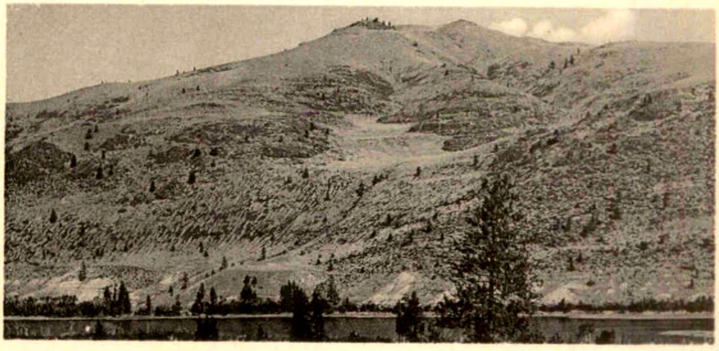

FIGURE 3. NORTH SIDE OF MCDONALD NARROWS
Denuded rock spur with deposit in shallow depression. Varved silt forms terrace at foot of slope.

HIGH EDDY DEPOSITS

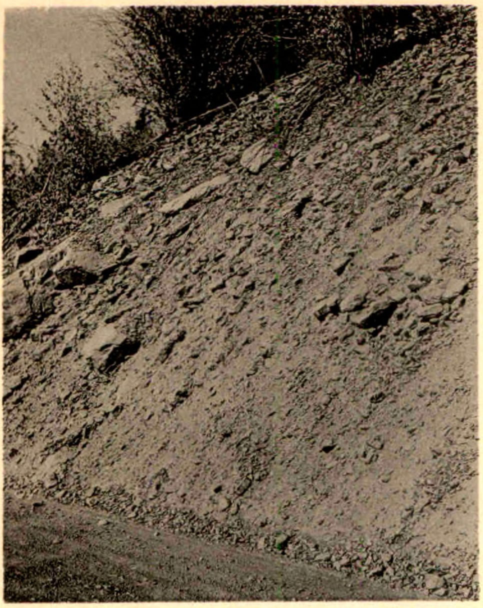

FIGURE 1. GRAVEL OF BOYER BENCH
Road grade up east front of terrace.

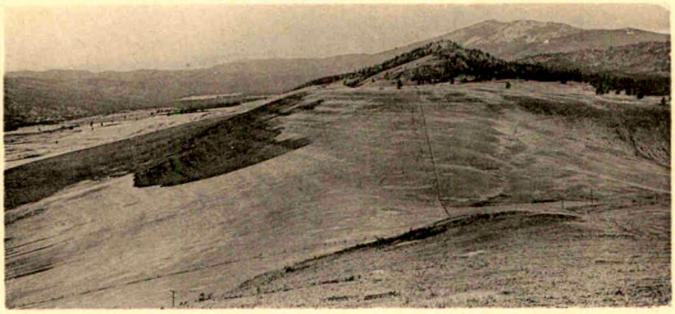

FIGURE 2. BEACHES ON WEST SIDE OF BOYER BENCH
Uppermost beaches turn into gap at right and end in small spitlike embayments.

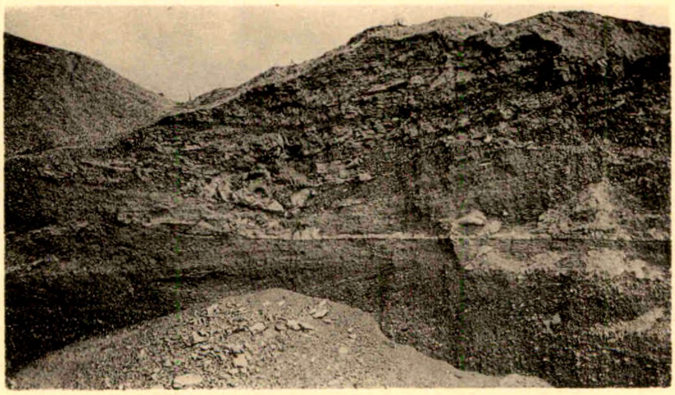

FIGURE 3. GRAVEL WITH FRAGMENTS OF VARVED SILT
Pit (Fig. 4, Sec. 9, T. 20 N., R. 26 W.) east of Clark Fork River near entrance to Eddy Narrows.

GRAVEL DEPOSITS AND BENCHES

of Knowles has been extensively cut away on the east side. In contrast, deposits that have been protected from stream cutting remain unmodified as illustrated by one in Quinn's Narrows (No. 21) that occupies a niche (Pl. 7, fig. 2) in the valley wall opposite Siegel Creek.

The gravel of the high eddy deposits consists of partly rounded pebbles mostly less than 2 inches in diameter in an abundant sandy matrix. In places the gravel is coarser, and a few large rough slabs like the rock of adjoining ledges usually are scattered through the mass. Gravel pits in deposits on Dunn Creek (No. 26) and on the west side of Boyer Creek Valley just above Plains expose distinct fore-set bedding.

At first glance the high eddy deposits suggest such features as deltas, bay-mouth bars, or terrace remnants of an alluvial valley fill. Attempts, however, to explain them as such meet with difficulties. Their back slopes descend in the opposite direction to the rising surface of a delta, and many of the larger ones are in pockets or disproportionately short gulches. Although a lagoonlike depression is generally present at the back the blockading ridge, unlike the horizontal top of a bay-mouth bar, has a sloping or rounded profile. The deposits, though situated in narrows where wave action was too weak to develop noticeable strand lines, are generally larger than the beaches and bars that were formed by the relatively strong waves in the wider basins. In addition their vertical distribution is without relation to any definite still stands of the lake. An interpretation of the deposits as remnants of a gravel fill of the river valley seems definitely disproved by their forms, vertical distribution, and the lack of correlative remnants of such an extensive deposit outside the narrows. On the other hand their diminutive counterparts may be observed in almost any rough-sided ditch or gutter after a flood has subsided.

Coarse gravel deposits in Clark Fork Valley.—Certain deposits of coarse bouldery gravel in Clark Fork Valley apparently are remnants of more extensive bodies laid down by a huge and rapid stream. One of these partly covered by sand and silt occupies a large area in the western part of Plains Basin. Its cross profile is slightly convex and along the east it has been partly cut away by the river from which it rises as a terrace 100 feet or more high (Pl. 1, fig. 1). Westward it extends to the side of the valley causing Swamp Creek to turn abruptly northward and hug the foot of the mountains (Fig. 4). The form and situation of this body suggests a channel deposit of a huge stream projected from Paradise Narrows, spreading and slackening somewhat as it crossed the basin to the outlet, the Eddy Narrows.

Where it enters the Clark Fork Valley just below the Eddy Narrows, the Thompson River has cut through a superficial layer of comparatively fine gravel (Fig. 2) and exposed a mass of large boulders over which it descends in a rapids. The position of the boulder mass suggests deposition by a huge stream rushing out of the Eddy Narrows.

About 4 miles below the Eddy Narrows a deposit of coarse gravel (Fig. 2), partly concealed by a superficial cover of fine sand, occupies a large area in the Clark Fork Valley adjoining Thompson Falls. From a terrace front facing the river it rises very gradually into a rounded barlike form that extends to the northeast side of the valley and turns the drainage from the mountains to a northward course along its margin. The form, position, and extent of this mass suggest a bar deposited by a stream large enough to have occupied the full width of the valley. Apparently the current turned against the southwest side from which it was deflected across the valley by a precipitous wall (Flatiron Ridge) below Thompson Falls, thus completing a wide bend. The current eroded the outside of the bend and deposited a bar on the inside. In the vicinity of Thompson Falls several large angular rocks lie on or in the gravel deposit. These may have been detached from ledges that crop out just above the town or transported from a distance by floating ice.

AREA AND VOLUME OF THE GLACIAL LAKE

The maximum altitude of the water surface in Glacial Lake Missoula is placed at 4200 feet as indicated by the uppermost of a series of beaches (Pardee, 1910). Repeated checks with the aneroid have indicated that the altitude of the highest definite beach is 4150 feet, although some doubtful beach gravel was found in one place (divide east of Oliver Gulch in T. 21 N., R. 22 W.) near the 4200-foot level. The outline of the lake at the 4150-foot level (Fig. 3) has been prepared with the aid of topographic maps and other surveys that were made after 1910. As thus determined, the area of the lake in round figures was 2900 square miles, and its volume 500 cubic miles. The areas, average depths, and volumes of water held by the different arms of the lake are roughly estimated (Table 2).

DEVELOPMENT OF THE UNUSUAL CURRENTS

Apparently the ice dam failed sufficiently to permit a sudden great outrush of water. The arm of the lake just above the dam was affected first, and there being no obstacle to its free discharge its surface fell rapidly throughout. Beyond Thompson River, however, the outflow was restricted by the Eddy Narrows and successively held back by the several narrows farther up. The lowered water surface in the arm above the dam induced a strong current through Eddy Narrows which lowered the water surface in the Plains Arm. As a result inflows were caused through Rainbow Lake Pass from Camas Prairie Arm and through the Paradise and Perma Narrows from that and other arms farther up (Fig. 9). The consequent lowering of the water surface in the Camas Prairie Arm caused an inflow across the submerged northern rim. In succession the other narrows released outflows from the more distant basins. The subsequent behavior of the currents thus generated was controlled largely by such topographic relations as the distribution and capacities of the different basins, the lengths and cross sections of the narrows, and in particular the "bottle necks" formed by the Paradise and Eddy narrows.

TABLE 2.—*Approximate areas, depths, and volumes of arms of Glacial Lake Missoula*
The water surface was at 4150 feet altitude.

Arm	Area (Square miles)	Average depth (Feet)	Volume (Cubic miles)
CLARK FORK VALLEY FROM MONTANA-IDAHO STATE LINE TO MOUTH OF FLATHEAD RIVER			
Lower Clark Fork (State line to Eddy Narrows)	465	1400	123.3
Thompson River Valley	115	600	13.0
Eddy Narrows	20	1380	5.2
Plains Basin	90	1350	23.0
Paradise Narrows	10	1330	2.5
FLATHEAD RIVER VALLEY			
Perma Narrows	10	1280	2.4
Camas Prairie Basin	95	870	15.6
McDonald Narrows	25	1200	5.7
Little Bitterroot Basin	260	900	44.3
Jocko Basin	105	900	17.9
Mission Basin	575	1100	120.0
CLARK FORK VALLEY ABOVE MOUTH OF FLATHEAD RIVER			
Quinn's Narrows	35	1200	8.0
Saint Regis River Valley	175	600	18.2
Superior Arm (Quinn's Narrows to Alberton Narrows)	120	940	21.4
Alberton Narrows	30	900	5.1
Missoula-Ninemile Basin	245	740	34.3
Bitterroot Basin	405	700	53.7
Hellgate Arm (Clark Fork Valley above Missoula Basin)	50	150	1.4
Blackfoot River Valley	65	440	5.4
Totals	2895 square miles		520.4 cubic miles

VOLUME AND DURATION OF THE UNUSUAL CURRENTS

Evidently the streams were of unusual size but were short-lived because of limited water supply. No precise data, of course, are available. The following results, based on formulas for small streams, are introduced merely to give some idea of their size.

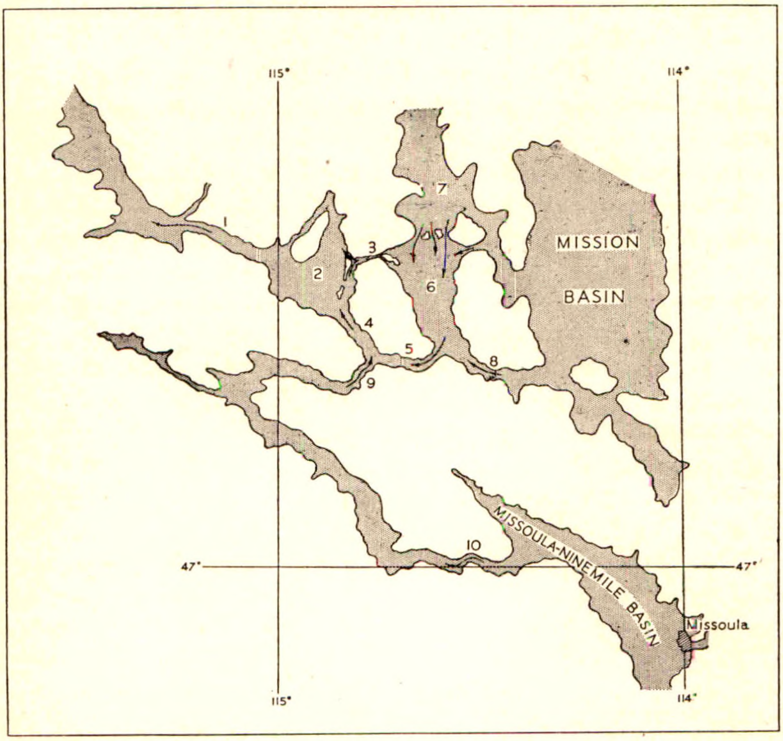

FIGURE 9.—*Central part of Glacial Lake Missoula*

Medium stage.

1. Eddy Narrows.
2. Plains Basin.
3. Rainbow Lake Pass.
4. Paradise Narrows.
5. Perma Narrows.
6. Camas Prairie Basin.
7. Little Bitterroot Basin.
8. McDonald Narrows.
9. Quinn's Narrows.
10. Alberton Narrows.

Eddy Narrows provided a nearly straight sluiceway 10 miles long with no very prominent projections along the sides to hinder the flow. As indicated by its high eddy deposits and the denudation of its sides the stream was at least 1000 feet deep. Available information of the upper limit is not precise enough to permit a satisfactory determination of the surface gradient, which is therefore assumed to have been the same as the average descent of the present river bed—namely, 2.3 feet per mile through the 10 mile stretch of the narrows. The average cross section is about 5,900,000 square feet, and the wetted perimeter 9600 feet. Using these figures with the Chezy formula (Merriman, 1931) for rough stone channels gives a velocity of 52.3 feet per second. According to the formula adopted by Manning (Lindquist, 1934) the same figures would indicate a velocity of 78 feet per second. The velocity of the eddying currents is rather closely related to that of the main stream (Leighly, 1934). As determined by a maximum pebble size in general of 3 inches (76 mm. approximately) the currents that formed the high eddy deposits had (Hjulstrom, 1935) a velocity of about 65 feet per second which may be assumed as about the same as the main current. Incidentally, this figure happens to be the mean of the two other determinations. Accepting it as approximately correct the flow through the Eddy Narrows would appear to have reached a peak of 386,500,000 second feet or about 9.46 cubic miles per hour. As only about 380 cubic miles of water was impounded above the Eddy Narrows the maximum flow could not have continued for long. The maximum discharge of the Mississippi River at Natchez in the flood of February 1937 (Williams and Crawford, 1939) was 2,100,000 second feet or about .05 cubic mile per hour.

Sufficient data are not available for even rough estimates of the flows carried by the other narrows or by the wind gaps around Camas Prairie or by Clark Fork Valley near the ice dam. However, according to data compiled by Hjulstrom (1935) the velocity of the stream that formed the giant ripple marks is indicated at one place (gravel pit in SE¼ sec. 12, T. 19 N., R. 24 W.) by a maximum pebble size of 40 mm. to have been 7 feet a second. This current presumably was at least twice as deep as the ripple marks are high, or about 100 feet, and it may have occupied the full width of the basin floor; obviously no estimate is possible without more data.

AGE OF THE UNUSUAL CURRENTS

Although they have remained exposed since the water subsided, the giant ripple marks and other features produced by the unusual currents show very little effect of weathering. Rock surfaces stripped of their surface mantle remain bare except for a relatively thin new cover. Talus aprons generally do not reach more than one third to half way up a cliff (Pl. 2, fig. 1). Except for later stream trenching in a few places, the giant ripple marks look as though they had just emerged. Where protected from stream erosion the high eddy deposits are likewise unmodified. Few of the many channel depressions in the wind gaps around Camas Prairie Basin have become filled. Large parts of the coarse gravel deposits in Plains Basin and Clark Fork Valley at Thompson Falls have survived attacks by the river. Altogether these conditions suggest an age not greater than Middle Pleistocene.

LAST STAGE OF THE GLACIAL LAKE

Either the rapid outflow was greatly slowed when the lake was partly drained or, if it had become emptied, the basin soon thereafter became, partly at least, refilled, and in this new body of water no unusual currents seem to have developed. The evidence for this delayed drainage or refilling consists mainly of beaches formed on the current-produced features of the preceding outflow.

A series of small beaches are plainly visible on the side of Boyer Bench facing Plains Basin (Pl. 8, fig. 2). From rather faint lines below Locust Hill they become more definite southward where the higher ones, as they turn eastward around the terrace, end in small spitlike embankments. Elsewhere rather faint strand lines are visible on a high eddy deposit along Vanderburg Creek east of Perma (NW¼ sec. 9, T. 18 N., R. 23 W., Fig. 5) and on the basinward side of the coarse gravel deposit in front of Wills Creek Pass (Pl. 3). A small gravel bar continuous with a beach at 3400 feet altitude lies across the lowest part of the stream channel of Markle Pass at its northern entrance. Another blocks the north entrance to a channel about 140 feet higher (Fig. 7). A similar relation between beaches and channel features exists at the northern entrance to Wills Creek Pass.

The highest of the later beaches is at 3540 feet. Whether that figure is the upper limit of the last submergence cannot be positively stated. Where the outflow features are absent present knowledge of the beaches is not sufficient to distinguish the younger from the older.

A relatively slow current during the final outflow may have formed the gravel bar in Plains Basin east of the river near the entrance to the Eddy Narrows. As exposed by a pit along the highway (Pl. 8, fig. 3) the gravel is fine-textured (about 15 to 20 mm., pebble size), clean, and its bedding fore-set. The gravel rests on a platform about 50 feet higher than the river and contains fragments of the varved silt.

SUMMARY

Failure of the ice dam in the Clark Fork Valley near the Idaho-Montana State line permitted the rapid draining of Glacial Lake Missoula estimated to have held 500 cubic miles of water. Rapid falling of the water surface caused huge currents in the narrows between submerged basins. Current-produced features include the unique giant ripple marks of Camas Prairie, a series of gravel ridges having the form, structure, and arrangement of ordinary ripple marks but of incomparable size. The flow through Eddy Narrows is estimated to have reached a maximum of nearly 9½ cubic miles an hour. After the outflow the basin again held a lake which drained away slowly.

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DEVELOPMENT OF THE NORTHERN ALLEGHENY SYNCLINORIUM AND ADJOINING REGIONS

BY G. MARSHALL KAY

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ABSTRACT

Analysis of the Paleozoic stratigraphy in the northern Allegheny synclinorium indicates that (1) Gross deformation was downward in early and middle Paleozoic. (2) The greatest stratigraphic thicknesses are in extensive semilenses of detrital sediment laid in depressions complementary to uplifts of the Taconian and Acadian orogenies, in deltageosynclines. (3) The Cincinnati and Adirondack axes geographically define the Allegheny belt; they had arches for limited times, were hinges of oscillation, and at other times were more depressed than the central part of the belt (4) The axes were less depressed than the elliptical Michigan geosyncline to the west and the linear Champlain geosyncline on the east (5) There may have been warping (epeirogenic) and sea-level (eustatic) movements synchronous with and related to the orogenies, as well as depression produced by loading (isostatic adjustment); the causes and interrelations are not proven, however

Later Paleozoic structures trend more southwesterly than the Adirondack axis; the rectilinear margin of the late Paleozoic (Virginian) geosynclinal belt crosses the axis in the central Appalachians The eastern margin of the southern part of the Allegheny belt is concealed beneath the first extensive Appalachian thrust sheet. Inasmuch as this sheet bears middle Ordovician rocks stratigraphically similar to those in the Champlain belt in the north, it forms an allochthone, and the thrust developed along the Adirondack axis Appalachian folds that extend into the northern Allegheny synclinorium parallel the Appalachian Structural Front, the western principal fold in the mountains. The Front has the arcuate trend of deltageosynclinal isopachs in the north and of the rectilinear trend of the margin of the Virginian belt in the south. The northern margin of the synclinorium was block-faulted in the early Tertiary.

INTRODUCTION AND ACKNOWLEDGMENTS

The Allegheny synclinorium is a broad, northeast-trending structural depression east of the Cincinnati arch (Fig. 1). Its northern margin extends to the faulted pre-Cambrian highlands of central Ontario and the Frontenac axis at the outlet of Lake Ontario. On the east lie the Adirondack block mountains. Farther south, the Appalachian Structural Front separates the deeply folded Appalachian Mountains from the gently warped border of the "geosyncline" beneath the Appalachian Plateau; though the front forms the margin of the present synclinorium, the extension of the Adirondack axis is farther east. The several limiting uplifts differ in character and times of origin, and the synclinorium is complex. Earlier deformation is evidenced in the distribution, thickness, facies, and conformability of the sediments, for much of the depression is of Paleozoic origin. Later history is recorded in the changed attitudes of the deposits.

The present survey has developed from stratigraphic studies of the Ordovician sediments along the northern and eastern margin of the region. Development of the region as a whole has been discussed rarely. Much information on which conclusions are based is in the literature, to which reference will be made continually. Additional data on subsurface have been furnished generously by the State surveys of New York, Michigan, and Ohio, and the Department of Mines of Ontario.

Students have continually encouraged the writer toward reasoned explanations of stratigraphic relations. The publications of Bucher, Grabau, Keith, Moore, Ruedemann, Schuchert, Stille, Ulrich, and Van der Gracht have particularly directed endeavors. W. O. Clif and Betty R. Smith have assisted in the preparation of illustrations.

The typescript was submitted originally in April 1940, as part of a paper, *Development of the northern Allegheny synclinorium and adjoining Canadian Shield;* the complementary part has been published as the *Ottawa—Bonnechere graben and Lake Ontario homocline* (Kay, 1942). The present article was awarded the George Frederick Kunz Prize in Mineralogy and Geology by the New York Academy of Sciences in 1941.

EARLIER PALEOZOIC STRATIGRAPHY

INTRODUCTORY STATEMENT

An understanding of the development of the Allegheny synclinorium must be gained from the stratigraphy and structure of the rocks of the region. The material presented in the following pages is principally descriptive, a summary of information accessible in scattered literature. It forms the basis for the interpretations in the latter part of the paper.

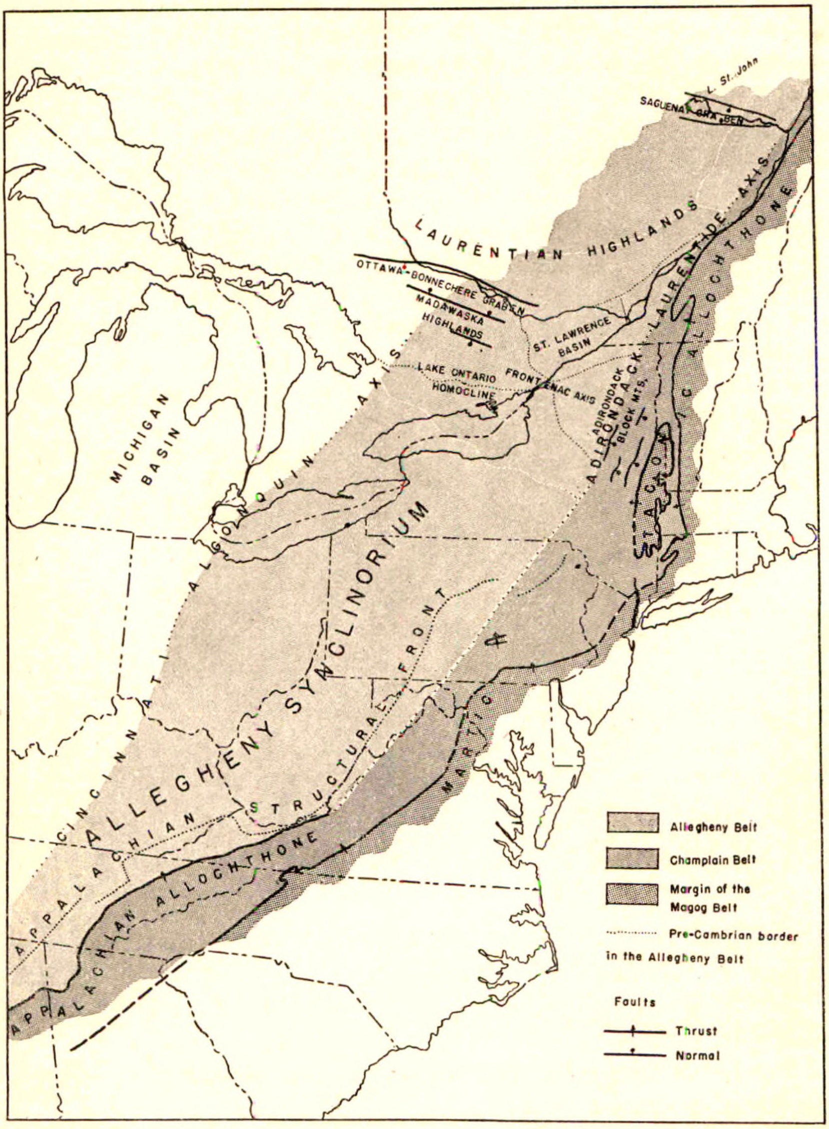


FIGURE 1.—*Map locating the Allegheny synclinorium, structural and stratigraphic provinces*

The limiting structures of the northern Allegheny synclinorium became defined in the Ordovician and Silurian. Throughout the earlier history of the belt, developments in the region to the east were particularly significant in its record. Belt and axis are to be used in a geographic sense without implication of structure. The Allegheny belt lies between the axis of the middle Ordovician Adirondack arch and the early Silurian Cincinnati arch. Although much of it lies within the Allegheny synclinorium, a present structure, the northeast part extends beyond the Frontenac axis, and the southern into the folded Appalachian Mountains and beneath the Appalachian allochthone.

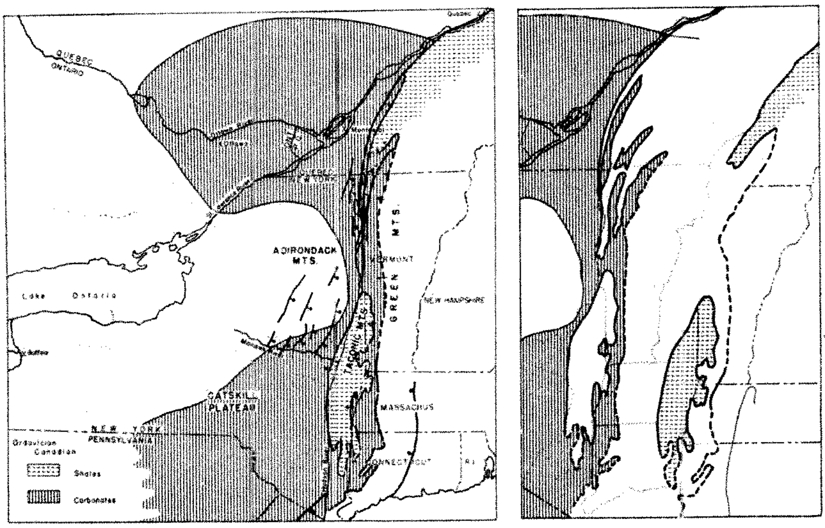
The Cambrian and earlier Ordovician rocks, sparse in the northern part of the synclinorium, are well represented in the Hudson-Champlain lowland east of the Adirondack Mountains. They were laid in the Champlain geosyncline, separated in the early Ordovician by uplifts along the Quebec axis from the Magog geosyncline farther east (Fig. 2); the axial structure directed and limited the westward extension of terrigenous lower Ordovician clastics, so that preponderantly argillaceous rocks of the Magog belt contrast with carbonates in the Champlain. The Adirondack arch came to define the western margin of the Champlain trough in middle Ordovician, separating it from the Allegheny belt.

In middle Trenton time, a belt of uplift, Vermontia, formed within the western part of the Magog geosyncline, resulting first in the introduction of terrigenous sediments in the Champlain trough and, in later Ordovician, into the Allegheny belt. At the close of the period, in the Taconian revolution, the argillaceous rocks of the Magog belt were thrust far westward, forming the Taconic thrust sheet that conceals the Quebec axis and all but the western part of the Champlain trough sediments.

The Adirondack and Laurentide Mountains lie along the Adirondack axis, and at least the former was block-faulted, probably at the beginning of the Silurian. Structure is first evident along the Cincinnati axis in early Silurian and continued to develop through the Paleozoic; the Michigan basin to the west also is clearly shown at this time. Lower Devonian Oriskanian sediments are limited northward south of Lake Ontario; the presence of equivalent rocks at Montreal implies a cross structure near the present Frontenac axis. Thus, the northern Allegheny synclinorium was delimited in the Devonian.

CAMBRIAN SYSTEM

The oldest sediments in the northern Allegheny belt are Upper Cambrian (St. Croixan). They overlap older Cambrian rocks of the Lake Champlain region and central Pennsylvania and extend to the vicinity of Utica on the Mohawk River, the Thousand Islands, and Arnprior on the Ottawa River (Fig. 3a); they are present in Adirondack outliers and northeast of Montreal. Though absent at Quebec City, where Mohawkian overlaps pre-Cambrian, they must be beneath the Taconic thrust plate to the south, for their boulders are present in the allochthonous lower Ordovician shales on the margin of the plate. Cambrian sediments are absent along the margin of the Paleozoic outcrop between the outlet of Lake Ontario and the head of Lake Huron but reappear on St. Joseph Island near the Michigan boundary. None are in wells penetrating to the pre-Cambrian southward to lakes Ontario and Erie or south of the former in New York. Questionably Cambrian rocks are present at depth in Sandusky County, Ohio, along the Cincinnati axis (Ballard, 1938, p. 1552). In Michigan, Cambrian is absent in the Detroit region and has not been reached by wells in the basin proper.



a. Present distribution of Canadian facies.

b. Canadian facies on palinspastic map.

FIGURE 2.—*Ordovician facies on present and palinspastic maps*

The carbonates in the Champlain belt were separated from the shales in the Magog belt by the Quebec "barrier". After Kay (1937) and Cady (1938).

The margin of Cambrian deposits is sinuous, a northwest-trending trough, the Ottawa Bay, extending from the Champlain geosyncline. Though the limit in the Adirondack region is not exactly determinate because outliers are few and poorly distributed, the data show the presence of a low peninsula lacking apparent correlation with the present mountains; the land Adirondackia (Schuchert, 1910, p. 463; Kay, 1937, p. 289) is probably misconceived. The westward thinning is best shown along the Mohawk Valley, where about 100 feet of Potsdam sandstone is overlapped westward by about 500 feet of Little Falls dolomite, which disappears in the West Canada Creek valley, probably through overlap and offlap (Ulrich and Cushing, 1910); there is no evident structural influence along the Adirondack axis.

ORDOVICIAN SYSTEM

Canadian and Chazyan series.—The limited distribution of Canadian and Chazyan rocks in the northern Allegheny belt is comparable to that of the Cambrian. At the beginning of the Ordovician, the Ottawa Bay continued to receive sediments, the distribution of which is summarized in Figures 3b and 3c.

Canadian rocks crop out westward to Little Falls on the Mohawk, the Thousand Islands on the St. Lawrence, and Allumette Island on the Ottawa, and trend northeastward from Montreal to north of Three Rivers. They offlap the Cambrian in New York and overlap somewhat in the Ottawa Valley. The thickness increases from a few tens of feet in Renfrew County, Ontario, to more than 2000 feet in the Lake Champlain region. Canadian rocks have not been encountered in wells penetrating to the pre-Cambrian in southwest Ontario or in New York and Ohio immediately south of the Great Lakes. They are present in the subsurface farther south in Ohio and, though absent at the Detroit River, have been found farther west in Lenawee and adjoining counties. A large peninsula occupying most of the Allegheny belt extended beyond the present lakes Ontario, Erie, and Huron.

That the Canadian history involved transgression, regression, lesser transgression, and withdrawal is shown by the relatively greater distribution of basal Canadian carbonates and the presence of younger Canadian disconformably on them, as at Ogdensburg, New York (Cushing, 1916).

The distribution of the Chazyan is somewhat similar (Fig. 3c). The overlapping beds thin northwestward from 1000 feet along the west margin of the Champlain trough; the thickness increases eastward in Vermont (Cady, 1938). The Chazyan overlaps in the Ottawa Bay, but the younger beds offlap northward along Lake Champlain (Raymond, 1906, p. 570). The most evident dissimilarity with distributions of older series is the absence of Chazyan south and southwest of the Adirondacks; and, whereas the Canadian and Chazyan are about 4000 feet in the Allegheny belt in central Pennsylvania, they are lacking farther north.

Mohawkian series.—Mohawkian sediments are extensive along the northern margin of the synclinorium and northeastward in the Allegheny belt. Their distribution shows a changed structural pattern from that of the Chazyan (Fig. 3d), evidencing the development of a foreland fold, the Adirondack arch, separating the Champlain trough from the newly formed Rideauan trough[1] in which the oldest Mohawkian of the region (Pamelia) was laid (Fig. 3d); the oldest Trenton beds (Selby member of the Rockland formation) have similar restriction (Fig. 5a). The Adirondack belt has thin and incomplete sections of the Black River and lower Trenton limestones; their irregular distribution suggests an undulating surface on which the seas had sinuous and embayed shores. The character of the belt is not conclusively shown, but there were no significant highlands. That it became an arch (Kay, 1936) separating the

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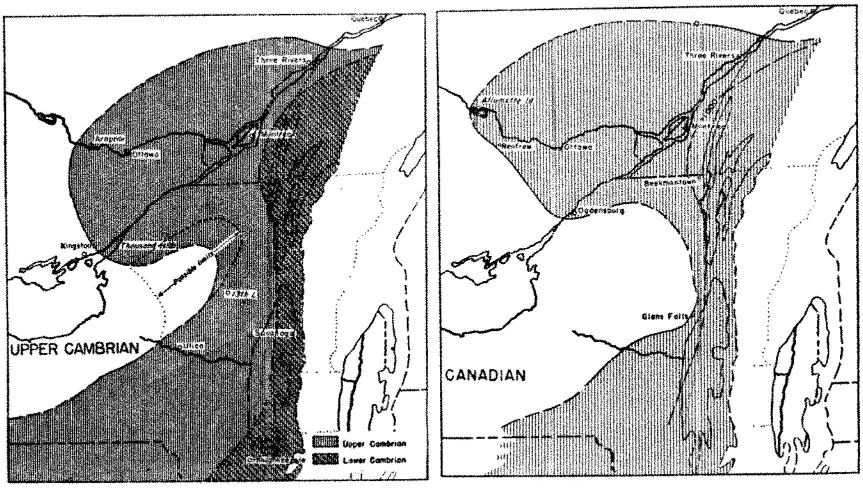
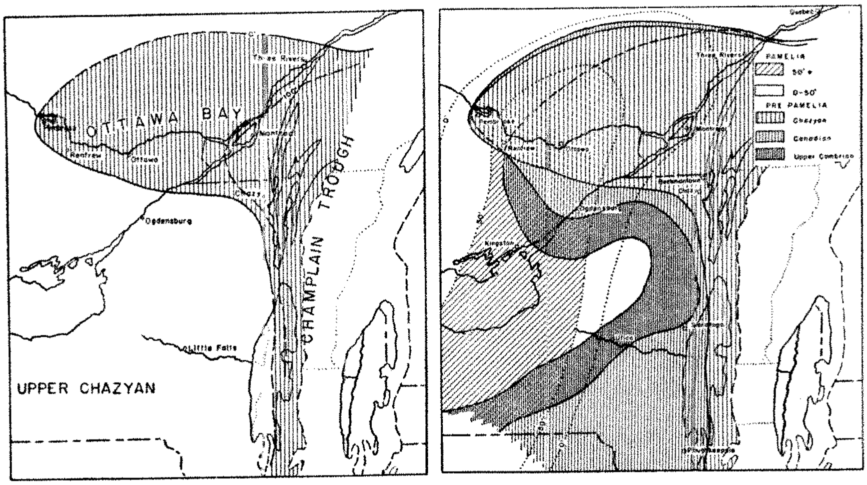
[1] Rideauan (new) replaces Ontarian (Kay, 1937, p. 289), the Rideau Lakes and River lying along the axis; Ontarian is a name too similar to Ontario basin (Schuchert, 1910, p. 473), applied to a different area, and is a homonym of the pre-Cambrian Ontarian geosyncline (Schuchert and Dunbar, 1933, p. 97).



a. Distribution of Cambrian series.

b. Distribution of Canadian series.



c. Distribution of Chazyan series.

d. Distribution of lowest Mohawkian (Pamelia) limestone on pre-Mohawkian geology.

FIGURE 3.—*Cambrian and Ordovician paleogeography*

Champlain trough from the depressed Allegheny belt is shown in the distribution and character of the Trenton formations.

The more restricted Mohawkian seas were margined on the north by shores trending east-west. The basal Trenton overlaps the Black River along a line extending from the north shore of Lake Huron toward Lake St. John, Quebec; and the higher Sherman Fall is overlapped by the Cobourg along a comparable line.

The Vermontian disturbance of middle Trenton time raised a highland within the western part of the Magog geosyncline and directly influenced the character and thickness of sediments in the Champlain trough, but terrigenous sediments entered the Allegheny belt only in latest Trenton (Collingwood). The disturbance is further reflected in the Allegheny belt only if the late middle Trenton marine withdrawal resulted from related eustatic fall or warping.

Mohawkian sediments overlap crystalline rocks over large areas (Fig. 3d), the Trenton extending beyond the Black River north of the present synclinorium. The oldest Black River (Pamelia) limestone evidences the southwestward trend of the axis of the Rideauan trough, being thickest at the northeast end of Lake Ontario and near Ottawa. The formation thins and overlaps toward Georgian Bay, is about 150 feet at the St. Lawrence River, and absent southeast of the Black River Valley. Along a parallel line northeast, it is absent at Mattawa, approaches 100 feet west of Ottawa (A. E. Wilson, 1934), thins eastward to Montreal (Okulitch, 1936) and northeastward (Okulitch, 1939) along the southeastern margin of the Allegheny belt. Pamelia sediments are lacking in the Lake Champlain region.

The Lowville is more extensive and is about 50 feet thick, north and east of Lake Ontario. It thins to extinction along the Mohawk toward the Adirondack axis and is relatively thin at Montreal and northeastward. It is present west of Lake Champlain but disappears east of the lake (Cady, 1938). The Lowville is overlapped by the Chaumont around pre-Cambrian inliers north of Lake Ontario and in exposures north of Lake Huron and in Lake Nipissing. Chaumont limestone is normally about 30 feet thick along Lake Ontario, thins northeastward, and is less extensive than the Lowville along the Mohawk and northern Lake Champlain. Thickness is normal at the southern end of the lake (Kay, 1937, p. 259), and Chaumont-like beds crop out in the outlier at Wells in the Adirondacks. Black River beds are absent at Quebec City and along Lake St. John (Raymond, 1916). Young (1937) has studied the stratigraphy of northwestern New York and southeastern Ontario.

In summary, the oldest Black River (Pamelia) is thickest in the Rideauan trough, thinning by overlap and convergence toward its margins. The Lowville is more extensive and in turn is overlapped along the northern margin of the trough by the Chaumont. Although the distribution in the Adirondack belt might be attributed to a breached arch, it is probable that the Lowville and Chaumont seas extended eastward as far as Vermont with embayed shore lines and that the beds had been somewhat eroded from the Adirondack axis in pre-middle Trenton.

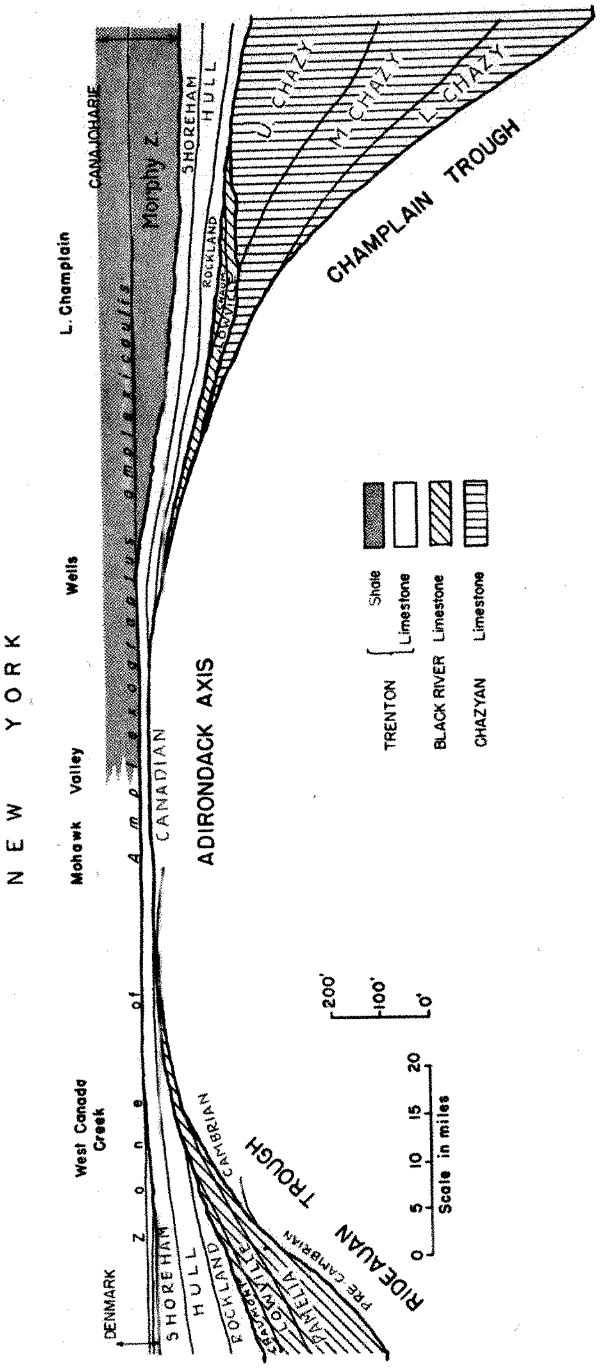


FIGURE 4.—*Restored section of pre-middle Trenton across the Adirondack axis in New York*

The principal representation of the Black River group is west of the Adirondack axis, the older beds being in the Rideauan trough. In contrast, the thick sections of the Chazyan are east of the axis in the Champlain trough, and the series enters the northern Allegheny belt only in the Ottawa Bay. Ulrich (1911, p. 414) considered these relations typical of oscillation; the explanation was somewhat confused by the misclassification of the Pamelia as inter-Chazyan but is fundamentally correct. There is oscillation along a section from Lake Ontario to southern Lake Champlain, in the sense that the Chazyan, thick in the latter area, is absent in the former, and conditions are reversed in the Black River (Fig. 4). There were epeirogenic movements, and the contrast continues southward in the Appalachian region. The Adirondack axis was a hinge, being tilted downward on the east and then on the west.

The lower Trenton Rockland and Hull formations are consistently present in the Allegheny belt and along the west side of Lake Champlain. Rockland seas were at first limited to the Rideauan trough, as evidenced in the distribution of the basal Selby member (Kay, 1937, p. 252-255; 1942, p. 598-604) (Fig. 5a), and the maximum thickness of the formation, about 70 feet, is northeast of Lake Ontario. If the Amsterdam limestone of the lower Mohawk Valley and southern Lake Champlain be lower Rockland (Kay, 1937, p. 259), there was a similar northeast advance in the Champlain trough. Younger Rockland overlaps the pre-Cambrian north of Lake Huron, Lake Nipissing, Renfrew County, Ontario, and toward Lake St. John, Quebec, where it lies on the crystalline rocks (Raymond, 1916; McGerrigle, 1933). There is similar overlap (Isle la Motte) on older Mohawkian and Chazyan sediments in the western Champlain belt. The distribution suggests that the Champlain and Rideauan troughs were depressed more than the Adirondack belt and that the latter was partially above the sea; the Adirondack arch had come into existence (Fig. 4).

The Hull is consistently present where post-Rockland sediments are preserved; 4 feet represented at Inghams Mills, New York, formerly was not recognized (Kay, 1937, p. 254). The formation has a maximum of about 100 feet in the Allegheny belt and approaches this in the Champlain. The beds overlap Rockland on pre-Cambrian monadnocks and are thought present in the basal Liskeard formation at Lake Timiskaming (Hume, 1925, p. 26). They are more extensive than the Rockland on the Mohawk, but absent on the Adirondack axis at Canajoharie, and are overlapped by the Shoreham near Quebec City on the Laurentide continuant of the axis. Inasmuch as Hull extended beyond the present outcrop, its limits can only be interpreted, but the isopachs reveal an arch on the Adirondack axis with greater depression on the two flanks.

The Shoreham limestone has been described as a member of the Sherman Fall formation (Kay, 1937, p. 264). The beds are mappable as far west as northwestern New York, and inasmuch as the principal stratigraphic contrast within the Trenton group succeeds them, they can be considered better as a third "lower Trenton" formation. As the Shoreham has not been mapped separately in Ontario, use of the term Sherman Fall continues west of New York; regional studies of thickness must be based on the larger unit. The Shoreham is the last Ordovician limestone east of the Adirondack axis. It thins and overlaps older Trenton on the axis and thickens westward into northwestern New York, reaching about 70 feet in southern Lewis County.

The similarity of Rockland and Hull faunas to those in Arctic America suggests their marine connection (Kay, 1935a, p. 588; Teichert, 1937, p. 44). The Timiskaming Strait (Kay, 1937, p. 289) is not tenable, however, for it is directed toward areas south of James Bay in which Ordovician and Silurian rocks are absent and probably never were deposited, inasmuch as the Silurian and Devonian overlap eastward toward the area (Dyer, 1929; 1930). The strait was conceived in the erroneous belief that the present relief was consequent on pre-Paleozoic elevations (Kay, 1942). If early Trenton seas did pass west of Ungava through western Hudson Bay, sediments of that age have not been recognized. The fauna of the Long Point series of western Newfoundland suggests the presence of lower Trenton (Schuchert and Dunbar, 1934, p. 72); intermigration from the Arctic to the Allegheny belt may have been along the east of Ungava.

The Vermontian disturbance of middle Trenton time raised a highland in the western part of the Magog geosyncline, and this directly influenced the character of the Denmark (Canajoharie) sediments in the Champlain trough. It is reflected in the Allegheny belt only if the marine withdrawal along the northern border resulted from eustatic fall related to the orogeny. That the Quebec axis did not form the place of the Vermontian uplift (Kay, 1937, p. 291) is shown in that west-marginal Magog trough deposits are still preserved in the boulder-bearing Levis shales along the St. Lawrence (Clark, 1923) and the Rysedorph conglomerate at Albany (Ruedemann, 1930), each in the Taconic allochthone; and early Denmark (Snake Hill) shales are present in both the allochthone and autochthone in the latter region (Ruedemann, 1930, p. 29, 117; Kay, 1937, p. 272). The Green-Sutton Mountain anticlinorium, within the Taconic allochthone, has been considered (Kay, 1940) the north limb of the structure responsible for the Vermontian clastics, the land Vermontia. The description of the position of Vermontia is complicated because later orogeny moved the belt in which it formed; Vermontia lay within the western part of the Magog belt, whose rocks have been moved westward so that the allochthonous position of Vermontia approximates the present position of the Green-Sutton axis and of the autochthonous Quebec axis. Later deformation is responsible for some structure of the present anticlinorium, however, there being infolded Silurian sediments along its eastern flank (Clark, 1934).

The disturbance had profound influence on sedimentation in the Champlain trough, which sank nearly a mile beneath the terrigenous sediments now preserved in the Hudson Valley. These shales and sandy shales are limited to the area on and east of the axis (Pl. 1). Relative depression along the sides of the axis is shown in the transgression of older formations by the basal Denmark (Fig. 4), beds older than the zone of *Amplexograptus* being limited to the two flanks, and in the presence of conglomerate-bearing Shoreham pebbles at the base of younger Denmark shales northeast of the Adirondack Mountains (Clark and McGerrigle, 1936).

The Sherman Fall (Shoreham plus Denmark) limestones are about 200 feet thick north and east of Lake Ontario but thin northward through the absence of younger beds. The formation is about 50 feet thick on the Escanaba River, Michigan, Manitoulin Island, and along the Ottawa, and about 100 feet northeast of Montreal; presumably only the Shoreham is represented in the thinner sections. The Cobourg limestone overlaps the Denmark north of a line trending eastward through southern Georgian Bay toward the northern boundary of New York; but the trend becomes northeastward along the Rideauan belt, for there seem to be thicker Sherman Fall beds north of the St. Lawrence than on the upper Ottawa (Parks, 1929). The reported 100 feet of Trenton at Lake St. John contains lower Trenton and Cobourg elements (McGerrigle, 1933); if present at all, the Sherman Fall must be quite thin.

The depth of water along the north shore of Lake Ontario can be judged, as the elevation of the beds is nearly 100 feet higher on the pre-Cambrian inliers than in the surrounding lowlands (Kay, 1942, p. 632). That the sea was shallower in the Allegheny belt than across the Adirondack axis is shown by pararipples common in Denmark calcite sandstones southeastward to the head of the Black River Valley but not present at Trenton Falls, where coarse interbeds are rare. Between this locality and Canajoharie, the Denmark grades through dense, lime-mudstones into thicker black shales of the Canajoharie formation. The Champlain trough was depressed deeply, permitting the settling of muds that would have remained in suspension had they reached the shallower waters on the west. The belt of gradation is west of the earlier structural arch, and the thickness of deposits and depth of water were greater across the axis than to the west; the Adirondack belt was one of flexure. The thickest clastics are in the Hudson Valley and southwestward into eastern Pennsylvania; the sediments are more calcareous along the Adirondack axis in northern New York than southwest of the mountains. The Vermontian highlands were most rapidly eroded and highest east of southeastern New York.

The principal contrasts in the Cobourg are the northward overlap of the limestone on the beveled Denmark in the Allegheny belt and the limited deposition of equivalent (lower and middle) Utica in the Champlain belt; the eastward continuity of the Utica black shale has not been demonstrated, and, though the Schenectady sandstone of the lower Mohawk has been correlated entirely with the Denmark, and Utica equivalents thought to be absent, the Utica possibly grades into the coarser facies southeastward. The limestones are about 150 feet thick, maintain this thickness northward in the Ottawa Valley, but diminish by offlap toward the Adirondack axis. The lower Cobourg limestone is fully developed as far southeast as near Poland, New York, with coarse calcite sandstones in the upper part; whereas the section above the lower Denmark at Middleville, 8 miles southeast, is entirely black shale, probably including beds equivalent to the Cobourg. There seems to have been a flexure or arch with its axis some 30 miles northwest of the axis of the lower Trenton Adirondack arch. The Cobourg extended across the Canadian Shield if it is represented in the Red River formation of Manitoba and its Arctic equivalents (Kay, 1935a, p. 586; Teichert, 1937, p. 43-44).

In latest Trenton (Collingwood-Gloucester), muddy sediments spread into the Allegheny belt and westward into the region of the present Michigan basin. This marked change in lithology has been interpreted as resulting from the filling of the Champlain trough, so that the terrigenous materials from Vermontia were carried into deepened waters transgressing the buried Adirondack arch. The relations initiated continued into the Cincinnatian epoch.

Subsurface Mohawkian.—In the Allegheny synclinorium and westward, the Black River and pre-Collingwood Trenton formations are limestones, contrasting with the argillaceous overlying beds and with the underlying crystallines and marginal sandstones. This lithologic unit is separated by the driller as the "Trenton lime." An isopachal map of these rocks must be based on a limited number of wells that have penetrated them, many wells being concentrated in small areas that yield oil and gas. The map (Fig. 5b) is limited eastward by the Adirondack belt in which the limestones grade into shales in the middle Trenton and are partly cut out by unconformity in the Cobourg.

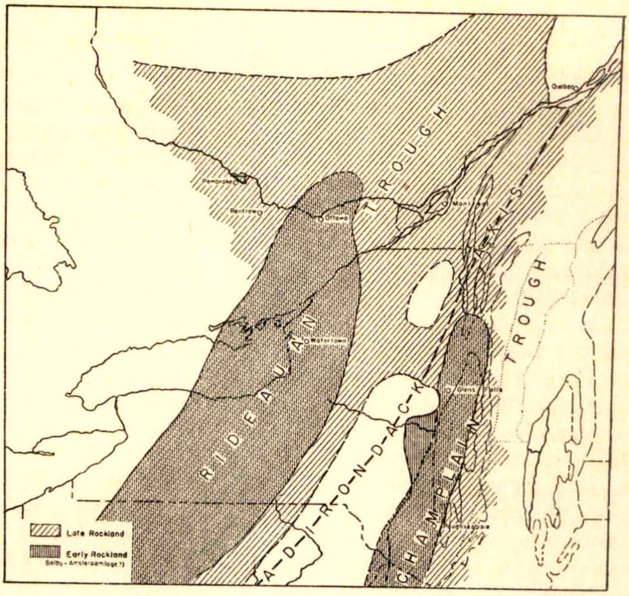
The first factor affecting thickness is the relief on the pre-Paleozoic surface. As a whole, this must have been quite small, for few wells give thicknesses varying widely from the average of many. Thus, of 15 wells in Kent and Essex counties, Ontario (Harkness, 1920-1936; Knight, 1915), the minimum is 790 feet, the maximum, 948; but only four vary more than 25 feet from the average of the whole number. The relief is well known on the outcrop, where partially exhumed monadnocks are scattered (Kay, 1942, p. 629-632).

A relatively thick belt extends from the Ottawa through Madison County, New York (Fig. 5b). The outcrop belts traversed are in the Rideauan trough in which the Pamelia is thickest; this factor accounts for most of the contrast with adjacent areas. A second trend of maximum thickness passes northwestward through Kent and Lambton counties, Ontario, lying along the Chatham sag in the Cincinnati arch, and evidences an early origin of this cross structure; this may be a belt in which the Black River is also thicker and more complete.

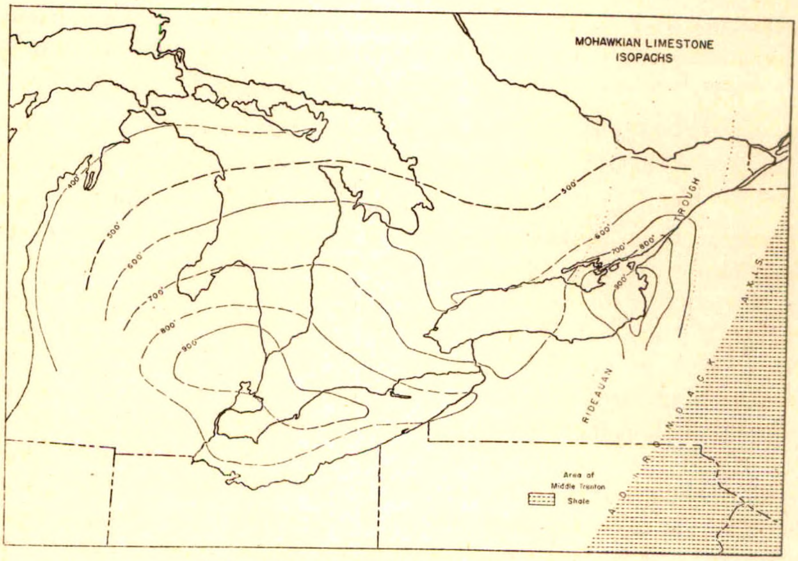
The thickness diminishes northward, due partially to overlap of the Black River group. Additional cause is the Sherman Fall-Cobourg disconformity, which results in the loss of some 150 feet of limestone between Lake Ontario and a line eastward from northern Lake Huron. There is some convergence in other formations, but it is not of great significance.

Thus the varying thickness of the Mohawkian "Trenton lime" reflects (a) irregularities on the pre-Ordovician surface, (b) the deposition of earlier rocks restricted to the Rideauan trough and possibly the Chatham trough with marginal overlap, and (c) the overlap of the Cobourg on diminishing Sherman Fall.

Cincinnatian series.—The sediments of the Cincinnatian in the Allegheny synclinorium are argillaceous and arenaceous, in contrast to the carbonates of the Mohawkian (Pl. 1). They are best exposed east of Lake Ontario (Ruedemann, 1925), are concealed westward beneath the lake, and sparsely shown along the outcrop north and east of the Niagara escarpment in New York and Ontario (Dyer, 1925; Parks, 1925; Fritz, 1926). Much of the understanding must be gained from logs of wells that have penetrated the series. The Cincinnatian converges northwestward (Fig. 6a), being 1000 feet along an arcuate line trending from the north end of Lake Simcoe to St. Thomas, Ontario, and from Sandusky toward Bellefontaine and Portsmouth, Ohio. The 1500-foot isopach lies about 50 miles eastward, near Hamilton and Port Dover, Ontario, Cleveland, Ohio, and in about western Roane County, West Virginia. The thickness exceeds 2000 feet in south-central New York and central Pennsylvania. The rocks form a low semi-cone of terrigenous sediments, having an apex in eastern Pennsylvania and evidencing a semilenticular depression of their floor.



a. Distribution of lowest Trenton (Rockland).



b. Isopachal map of Mohawkian limestones.

FIGURE 5.—*Mohawkian paleogeography*

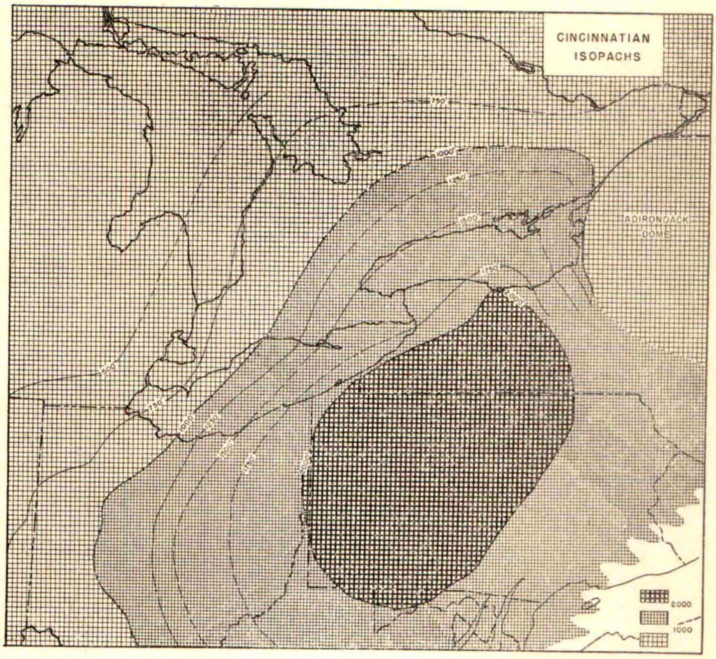
The series has two principal facies in the northern part of the synclinorium—the lower gray shales and sandy shales, and the upper red shales (Queenston) of Ontario and Ohio that pass into coarser red beds (Juniata) southeastward. The red facies gradationally overlaps late Cincinnatian (Richmond) gray shales northwestward and is overlapped by Silurian conglomerates southwest of the Adirondack Mountains and in northeastern Pennsylvania. Although the sub-Richmond Lorraine or Eden-Maysville converges northwestward, the addition of lower Richmond shales of similar lithology gives the facies a fairly constant thickness of about 1000 feet in New York, Ontario, and northern Ohio. The Lorraine retains this thickness to the limit of its outcrop west of the Adirondack Mountains (Kay, 1939); it probably continued over the Adirondack axis without significant change, possibly with expansion, and across the Champlain trough to the source in Vermontia.

There was late Maysville uplift southeast of central and eastern Pennsylvania, for the sandstone and conglomerate of the former area exceed 1300 feet. Thin conglomerate at the base of the Shawangunk (Silurian) conglomerate that has been correlated lithologically with the Bald Eagle (Maysville) conglomerate lies unconformably on the older Lorraine in eastern Pennsylvania (Willard and Cleaves, 1939); the correlation seems tenuous. The effect of the disturbance is reflected in the coarse texture of the uppermost Lorraine, Oswego sandstone, which spread as far as central New York and northern Virginia. The disturbance did not extend far northward as shown by the absence of sandy facies in the thick section on the Nicolet River south of Three Rivers, Quebec (Foerste, 1916, p. 14-41). Moreover, the great thickness of the Lorraine in that area shows continued depression of the northern Champlain trough.

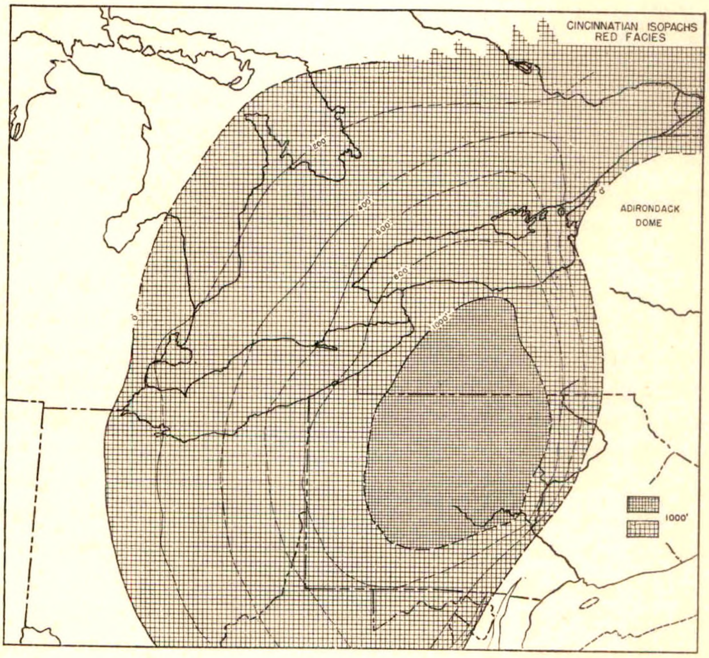
The Queenston red facies (Fig. 6b), the base rising stratigraphically northwestward by gradational overlap, extends to eastern Ontario, Lake Huron, southeastern Michigan, central Ohio, and southern Virginia with an arcuate margin; gray shales with marine faunas comprise the whole of the Cincinnatian peripherally. The Richmond shales overlap the Lorraine between Manitoulin Island, where both Eden and Maysville are present (Caley, 1936), and northern Michigan (Hussey, 1926). The Richmond (Juniata) red beds are absent in eastern Pennsylvania (Willard and Cleaves, 1939), extending only into the western margin of the Champlain belt, and similar relations pertain southwestward into Virginia (Butts, 1933). The equivalent Queenston is overlapped by Silurian southwest of the Adirondack Mountains.

The area of the Adirondacks and of eastern Pennsylvania lay to the east of the graded plains of the sinking basin in the Allegheny Belt. In southeastern Quebec and westward through Ontario, marine waters persisted through early Richmond (Waynesville); and more than 1000 feet of fine-textured Queenston in Quebec was laid in the sinking Champlain trough. The seas retreated before the advancing terrigenous sediments, producing the gradational overlap of the red shales over the gray marine shales. Isopachs of the red facies (Fig. 6b) do not represent synchronous deposits but support the view that the principal depression was toward eastern Pennsylvania.

The Cincinnatian isopachs cross those of the Trenton in western Ontario. There is no appreciable abnormal thinning of the sediments on



a. Isopachal map of Cincinnatian series.



b. Isopachal map of red facies of Cincinnatian.

FIGURE 6.—*Cincinnatian isopachal maps*

the east flank of the Cincinnati axis, and the facies lines do not evidence influence of an arch or flexure. The Cincinnati arch is believed of later origin.

TACONIAN REVOLUTION

The Taconian revolution[2] at the close of the period involved thrusting of rocks deposited in the Magog geosyncline on the folded strata of the Champlain trough; the overthrust rocks form the Taconic allochthone (Figs. 1, 7). The principal area of allochthone is the Taconic klippe, composed of argillaceous Cambrian to middle Ordovician lying as a tectonic outlier of stratigraphically exotic rocks surrounded by equivalent carbonates of more westerly origin in eastern New York, southwestern Vermont, and western Massachusetts. Similar "outliers" have been demonstrated allochthonous in western Newfoundland (Johnson, 1940); the anomalous character of the facies and faunas suggests this relationship, though the beds were considered older than the carbonates previously (Schuchert and Dunbar, 1934, p. 38-40). The Taconic thrust evidently lies along White Bay, for the sediments on its west (Betz, 1939) are of Champlain trough facies, those to the east in Notre Dame Bay (Heyl, 1937) of the Magog. In Pennsylvania, similar exotic shales have been interpreted as possibly in a klippe of the Martic thrust sheet (Kay, 1941a). The minimum displacement, disregarding the effects of retreat of its most advanced front through erosion, foreshortening in later movements (Cloos, 1940), and the gap that must have separated the westernmost preserved allochthone and easternmost exposed autochthone, was about 40 miles in Newfoundland and in New York-Massachusetts, (Kay, 1937, pl. 5; Schuchert, 1937, p. 1028), and 30 miles in Pennsylvania. The true displacement was considerably more.

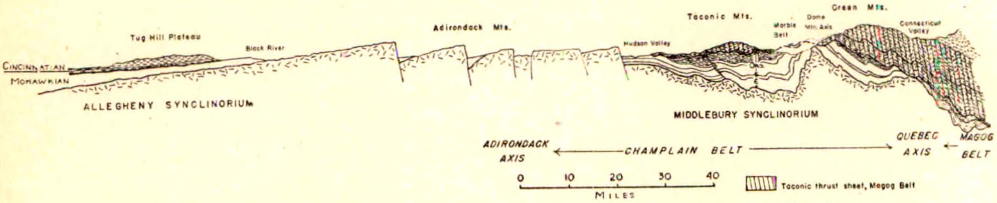


FIGURE 7.—*Structure section from Lake Ontario to New England*

Inasmuch as the marginal allochthonous sediments are of Magog belt facies, the fault was initiated within the border of that geosyncline. That it was adjacent to the Quebec axis is shown by conglomerates with exotic

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[2] The writer and others have referred to the post-Ordovician orogeny as the Taconic disturbance (Kay, 1937, p. 291; 1940), a term less desirable than Taconian revolution of other authors; the use of Taconic for both a thrust sheet and an orogeny leads to confusion, and the lateral movement, one of the greatest in North American history, warrants the term revolution.

(Champlain belt) boulders in Quebec (Clark, 1924) and New York (Ruedemann, 1901; 1930). In Maryland, the Champlain belt sediments near the Martic thrust thin and diminish in content, as though by overlap (Stose and Jonas, 1936).

The thrust sediments are in tectonic contact on Queenston shale in southeastern Quebec, and the autochthonous Cincinnatian has been folded considerably (Parks, 1931, p. 29-30). The overthrust rocks are overlain at Becraft Mountain, New York (Kay, 1937, p. 287-288) and in the Catskill Front (Ruedemann, 1930, map) by latest Silurian Manlius limestone. Thus, there is direct evidence that the principal lateral movements were pre-Manlius and post-Queenston. Folds in autochthonous Ordovician are truncated by the Shawangunk and Tuscarora quartzites of the earliest Silurian in southeastern New York and Pennsylvania; if the folding accompanied Taconic thrusting, the revolution is pre-Silurian.

The front of the thrust sheet was not very high. Middle Ordovician sediments are preserved near to the westernmost remnant of the sheet and probably never were buried deeply. On Anticosti Island in the Gulf of St. Lawrence, there is essentially continuous section of Cincinnatian and early Silurian calcareous shale and limestone (Twenhofel, 1927) in the Champlain belt within 50 miles of the overthrust rocks of Gaspe; the allochthone was beneath the sea or not high enough to produce significant detritus after the revolution. Though the quantity of Silurian terrigenous sediments is distinctly smaller than that of the Ordovician, as will be shown later, this reflects repeated uplift and continued presence of Vermontian highlands in later Ordovician, in contrast to progressive reduction of the transposed Taconia in the Silurian. The greatest quantity of eroded material was laid in the latitude of Pennsylvania, as shown by isopachs; that the greatest elevation was there also is shown by the coarser texture of the sediments. The lateral movement of the allochthone may have been as great or even greater in Quebec, but Vermontia and its transposed descendant, Taconica, were more continually high farther south. Quantity of erosion is measured by volume of sediments; elevation and proximity of source by texture.

The metamorphic pre-Cambrian rocks on the sole of the thrust where it is exposed farthest east in Pennsylvania probably show preceding uplift in the core of the Vermontian geanticline. It has been suggested that the thrust passes west of the Green Mountain anticlinorium east of Rutland, Vermont (Hawkes, 1941); the metamorphic rocks may correspond to the Glenarm sediments of Manhattan and southwestward. The presence of later Ordovician slates from Pennsylvania to Virginia in the Martic thrust block implies that the axis of Vermontia lay west or east of the metamorphics adjoining the fault, and their structure evidences post-Vermontian compression. The hundreds of cubic miles of detritus that came from Vermontia and Taconica demand raising of a considerable belt within the Magog geosyncline; the metamorphics would represent what lay beneath the floor of the Paleozoic geosyncline. If Ordovician rocks are to be found, they should be farther southeast and correspond to those of the New Hampshire synclinorium (Billings, 1936) whose western limb is within the Magog belt.

"Appalachia" is a term applied to undefined land southeast of the Appalachian region sediments, just as "Llanoria" is utilized in south-central United States (Van der Gracht, 1931a, p. 118) and "Cascadia" in the west. Referring to the rocks of the Martic anticlinorium—"the embryonic ranges which formed as a result of the first post-Cambrian orogenic deformation arose in the eastern half of the Ordovician geosyncline" (Bucher, 1933, p. 132)—Vermontia developed in the Magog geosyncline, within the Appalachian belts. The Glenarm was laid in a pre-Cambrian geosyncline, evidently the precursor of the Magog, but it was not deformation at the close of the Cambrian (Jonas, 1929, p. 510) that elevated the rocks, but the Vermontian disturbance in the Ordovician; one must not infer that the Martic anticlinorium is the Vermontian geanticline, for it was later deformed.

SILURIAN SYSTEM

Niagaran series.—The earlier Silurian Niagaran series is separable into the Alexandrian, Clinton, and Lockport groups, each present in the synclinorium (Pl. 1). The inclusion of all pre-Cayugan Silurian in a single series (Grabau, 1921, p. 314) seems advantageous; the typical rocks formed with essentially progressive overlap, though incomplete at the base and within in New York State. Prior to the Taconian revolution, the region of the present Adirondacks formed a broad dome in late Cincinnatian. Following the orogeny faulting along the east flank produced the present block-mountain structure. The time of faulting has not been conclusively established, but is considered early Silurian (Megathlin, 1939, p. 119-120); what seemed most significant, the considerable displacement on the Mohawk and its seeming absence in the Silurian of the Helderberg Plateau near by (Kay, 1937, p. 288) has been found due to termination of the faults rather than their being overlapped (Megathlin, 1939). On the northwest side of the synclinorium, the Alexandrian and particularly the Clinton evidence establishment of an arch along the Cincinnati axis. Thus the western limit of the Allegheny belt had become established in early Silurian.

The Albion or higher Alexandrian has been studied thoroughly along the Niagara escarpment in Ontario (Williams, 1915a; Cumings, 1939) and in wells south and southwest of the outcrop (Harkness, 1928; Newland and Hartnagel, 1932, 1936; Fettke, 1935; Hartnagel, 1939). The beds change from dominantly sandstone in north-central New York to limestone and intercalated shale in the Michigan basin (Pl. 1). The sandstones are limited to the southeast of the present Cincinnati arch, practically disappearing northwest of a line from Port Stanley through Brantford toward Toronto (Harkness, 1929, p. 67-68); they are similarly restricted to the east flank of the arch in Ohio (Ballard, 1938, p. 1534). Seas passed over the axis and are represented principally by carbonates in the Michigan basin. North of the extension of the axis, at Lake Timiskaming, beds have facies and faunal characters of the Albion of the Michigan basin (Hume, 1925, p. 28-33).

Eastward in New York, the Albion is overlapped by the Clinton, the Oneida conglomerate (Sanford, 1936, p. 803-804) forming the marginal facies. The sandstones retain thickness of about 100 feet through western New York, approaching 200 feet in northern Pennsylvania wells. In central Pennsylvania, the Tuscarora quartzite is nearly 1000 feet; only the upper, Castanea, member is thought equivalent to the New York Albion (Swartz, 1934, p. 111). The older Alexandrian sediments must converge and overlap rapidly in north-central Pennsylvania. Eastward in Pennsylvania and in southeastern New York, the Alexandrian is contained in the lower part of the thick, conglomeratic Shawangunk formation, separated from the Albion of central New York by the south-plunging Adirondack arch. The arch shielded the northern Allegheny belt from direct influx of sediments from the highlands of Taconica. Rapid erosion is implied by the coarse texture and quartz content of the detritus. The trough in the Allegheny belt plunged gently southward in southern New York, the floor deepening rapidly below the early Silurian semilenticular mass in eastern Pennsylvania.

Clinton sedimentation simulated that of the Albion, but the group is absent along the Cincinnati arch in Ontario east of Lake Huron, and a contrast in facies is represented on the flanks. The Clinton disappears across the "Cataract Axis" (Fig. 4) (Schuchert, 1911; Williams, 1915b) along the trend of the Cincinnati arch, the structure evidently plunging southwest toward the Chatham sag, where the Clinton breaches the axis. The margin of Clinton shales runs from St. Marys through Stratford and Kitchener (Harkness, 1929, p. 66). Northwest of the peninsular arch, in the Michigan basin limestones extend southeastward to the south end of Georgian Bay (Cumings, 1939, p. 596). Though Clinton is not reported at Lake Timiskaming, beds may have been included in the lower part of the "Lockport", as has been done in the past in Michigan.

Thus, the Cincinnati arch is first clearly indicated in the Clinton. This advent is about synchronous with that in Ohio, where Albion sediments persist but are thin on the arch, whereas the Clinton offlaps (Foerste, 1935). That the arch was not present in the Ordovician has been indicated from the outcrop and isopachs; the Chatham sag was suggested by Mohawkian isopachs, however. The evidence for post-Ordovician age in Ohio and Kentucky has been summarized by McFarlan (1939); structural trends during pre-Mohawkian and Mohawkian time are across the trend (McFarlan, 1938). Moreover, the Cincinnatian is quite fully developed on the axis where the lower Silurian is poor and incomplete.

Clinton sediments in New York have lensing section, thicken eastward to Onondaga County, and change to coarser facies eastward, the latter, the Oneida, disappearing in Herkimer County (Chadwick, 1915; Sanford, 1936). There is northward convergence from 800 feet in central Pennsylvania (Swartz, 1936) to 200 feet in western New York and about 300 feet in Onondaga County. The character of the Oneida, with white quartz pebbles, implies an origin similar to the partially equivalent Shawangunk; the pebbles evidently came around the south flank of the Adirondack arch.

The Lockport group is persistently carbonate and rather constantly 150 to 250 feet thick, from Georgian Bay to western New York, thinning and disappearing by offlap in Onondaga County. The lower Lockport (Manistique) is overlapped southeastward from the Michigan basin (Cumings, 1939), the basal dolomite in southern Ontario and New York being the Engadine, lying disconformably on Clinton. The beds become argillaceous south of the deepest wells in central New York if the McKenzie shaly limestone of central Pennsylvania corresponds (Swartz, 1934), and somewhat thicker; it has not been determined whether these include pre-Engadine Lockport and are basally overlapped northward. The McKenzie passes eastward into the lowest red shales of the Bloomsburg of eastern Pennsylvania (Swartz and Swartz, 1931).

The Adirondack belt in New York lacks deposits older than latest Silurian; evidently it was low, for the Lockport is not terrigenous to the west. Lockport blocks have not been found in the volcanic breccia at Montreal; possibly there was an arch as to the south of the Adirondack Mountains.

Cayugan series.—Cayugan rocks are exposed in three isolated areas: along the southeast of the Michigan Basin; north of eastern Lake Erie and south of Lake Ontario along the northern margin of the Allegheny synclinorium; and in the folded Appalachians of Pennsylvania, Maryland, and the Virginias. Exact correlations are uncertain, but the gross deformation can be ascertained, for wells penetrate the series in southern New York, eastern Ohio, and southwestern Ontario (Pl. 2b). The highest Cayugan Manlius limestone of New York seems equivalent to the Keyser limestone of the Appalachians (Swartz, 1931; 1939); these beds are absent in western New York and southern Ontario but are thought equivalent to the upper Bass Island group in western Ontario (Williams, 1915, p. 93-94).

From Erie and northern Chautauqua counties, western New York, to Elgin and southern Oxford counties, Ontario, the series is composed of 350 to 400 feet of gypsum-bearing dolomites (Fig. 8). Westward, the rocks thicken to more than 1000 feet near Detroit, and exceed 2000 feet in Livingston County, Michigan. Part of the thickening is by addition of higher beds, but it is considerably due to divergence, an indication of the sinking character of the Michigan basin region and the adjoining Cincinnati axis. Eastward in New York, it thickens to more than 1000 feet in Onondaga County, then thins, and the Manlius overlaps on pre-Silurian beds east of Herkimer County over the Adirondack axis. Along this section, the basal Cayugan Vernon red shale enters in Genesee County, exceeds 500 feet in Onondaga County, then disappears eastward; it is succeeded by the principal salt beds.

The isopachal map (Pl. 2b) shows a southwestward-plunging basin in New York west of the Adirondack axis, the Cayugan thickening to more than 2000 feet in eastern Schuyler County, where the lower 800 feet is of Vernon red and green shale (Hartnagel, 1939). More than 2800 feet of Cayugan has been penetrated in Tioga County, northern Pennsylvania, but the total decreases westward to less than 1000 feet in southern Cattaraugas and Chatauqua counties, southwestern New York. The maxima then trend southeastward into eastern Pennsylvania, where nearly the whole section is Bloomsburg red shale, exceeding 2000 feet, with succeeding upper Tonoloway and Manlius limestones (Swartz and Swartz, 1931); in central Pennsylvania there is about 1200 feet of Cayugan (Butts and Moore, 1738; Butts, 1939).

There is no information on subsurface Cayugan through western Pennsylvania. Across West Virginia, the Albion to Oriskany interval, including beds additional to Cayugan, decreases from 2400 feet in Randolph County to less than 1600 feet in Roane, and 1435 in Wood County (Martens, 1939), suggesting convergence of the Cayugan that comprises most of the unit. Increase to 1700 feet in Harrison County, Ohio (Ballard, 1938), suggests that the Allegheny belt was a broad arch separating the Champlain belt from the Michigan basin as in the north.

The tongues of lower Cayugan red shales extend southwestward into northern Virginia in the base of the Wills Creek, and in New York to Genesee County; their limit is arcuate. Saline rocks lie northwest of the lower Cayugan red shale semicone; southward the limy shales of the Wills Creek are marine. Flow of marine water to the interior was restricted by the Bloomsburg-Vernon red shale delta and the relatively shallowly flooded Allegheny belt in West Virginia, Pennsylvania, and southern Ontario, isolating the northwestern Allegheny belt and Michigan; precipitates formed in these partially enclosed, sinking interior areas. The two saline basins were partially separated along the middle of the Allegheny belt. The basin extended northwestward to the west of Ungava, for there are dolomitic shales southwest of Hudson Bay (Dyer, 1930).

The youngest Cayugan limestones overlap the Adirondack axis in eastern New York and extend eastward over the margin of the Taconic allochthone. They are present in the eastern border of the Allegheny belt, and in the Champlain belt, and may be represented in the upper Bass Island carbonates of the Michigan basin. The Allegheny belt was less depressed than the adjoining geosynclines.

There is much less terrigenous material in the post-Taconian Silurian "molasse" than was derived from Vermontia in the later Mohawkian

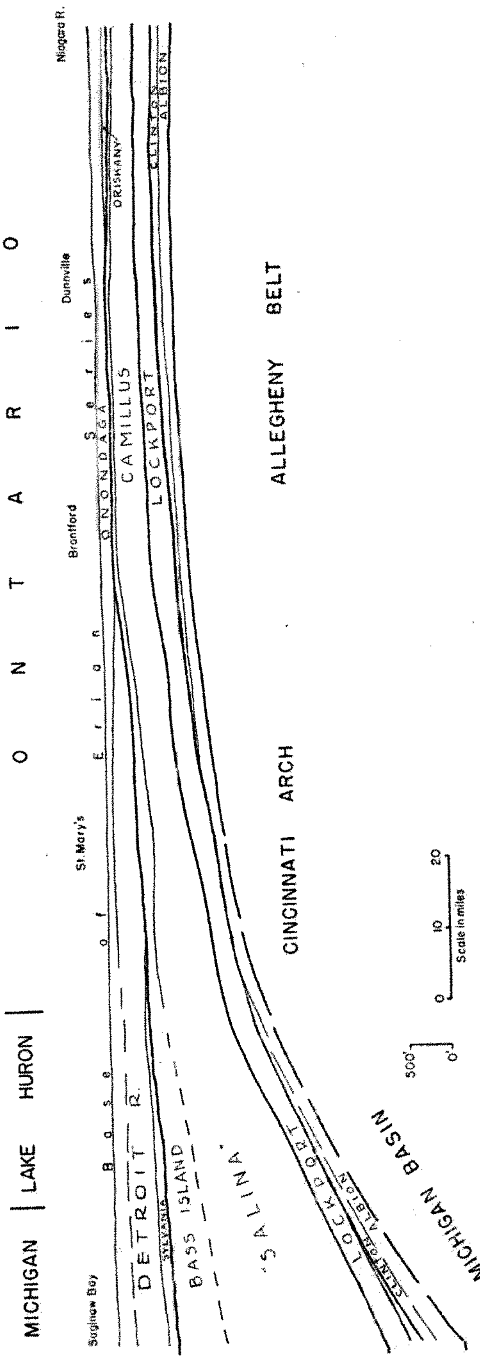

FIGURE 8.—*Restored section of Silurian and older Devonian across the Cincinnati axis in Ontario*

and Cincinnatian (Fig. 6a; Pl. 2a). Within the area of preservation east of the Cincinnati axis in Ohio, Pennsylvania, Ontario, and New York, the quantity of the former approximates 65,000 cubic miles, of the latter, 95,000. The Ordovician quantity should be almost doubled, for it excludes a great volume of Mohawkian and Lorraine in the Champlain belt and a smaller amount of Cincinnatian shale west of the Cincinnati axis.

DEVONIAN SYSTEM

Helderbergian, Oriskanian, and Ulsterian series.—The Lower Devonian Helderbergian and Oriskanian series are restricted to the area east of the Cincinnati axis. The latter are the oldest rocks penetrated by many wells within the synclinorium in Pennsylvania. The greatest depression was in a trough having more northeasterly trend than the Ordovician Adirondack axis, and the margin of the sea was within the western Champlain belt; the structural influence that directed the form and trend of the Adirondack axis had been superceded. The Oriskanian evidences the relatively higher structure of the Cincinnati belt and the separation of the Allegheny basin from the St. Lawrence, possibly along the Frontenac axis.

Helderbergian limestones were laid disconformally on the Manlius in eastern New York and Pennsylvania. They continue across the Adirondack axis into the Champlain belt with eastward thickening by offlap in New York and are absent in central and western New York. They are nonterrigenous east of the Hudson at Becraft Mountain and represented by limestone blocks in the St. Helens volcanic breccia at Montreal, signifying that the Adirondack area and Taconic allochthone were rather low. The Allegheny belt was tilted southeastward; the axis of the New Scotland (Swartz, 1939, p. 57) and earlier troughs cross the Adirondack axis in Pennsylvania. The limestones pass into marginal sandstones in western central Virginia (Swartz, 1929), within the Champlain belt.

The margin of the Oriskanian is rather clearly delimited in southern New York and northwestern Pennsylvania (Fig. 9) (Fettke, 1938, p. 241; Lafferty, 1938, p. 175; Martens, 1939, p. 30-35) and along the northwest; it paralleled the Cincinnati axis, approaching most closely at the Chatham sag. The sands were laid in a shallow, southward-deepening sea and are locally absent within the synclinorium, as in southern McKean County, Pennsylvania (Fettke, 1935, Fig. 1). They extend eastward into the Champlain belt in New York and Pennsylvania (Cleaves, 1939). Oriskanian blocks in the St. Helens breccia at Montreal (Williams, 1910) indicate that the Adirondack region had become connected to the shield; the margin of the sea that trended eastward south of Lake Ontario evidently turned northward around the east flank of the Adirondacks and was embayed in the upper St. Lawrence basin. The trend and character of the Devonian arch cannot be delimited but may have been in the position of the Frontenac axis.

The Ulsterian is more extensive, the Onondaga limestone in the Allegheny synclinorium and the Detroit River series of the Michigan basin being representatives (Warthin, 1939, p. 601-602). It forms the last dominantly carbonate sediment in the region, the advent of Acadian orogenic movements in the Erian leading to the formation of westward-thinning terrigenous sediments of the later Devonian (Cooper, 1931; Reeves and Davies, 1937). The Onondaga persists with a thickness of 100 to

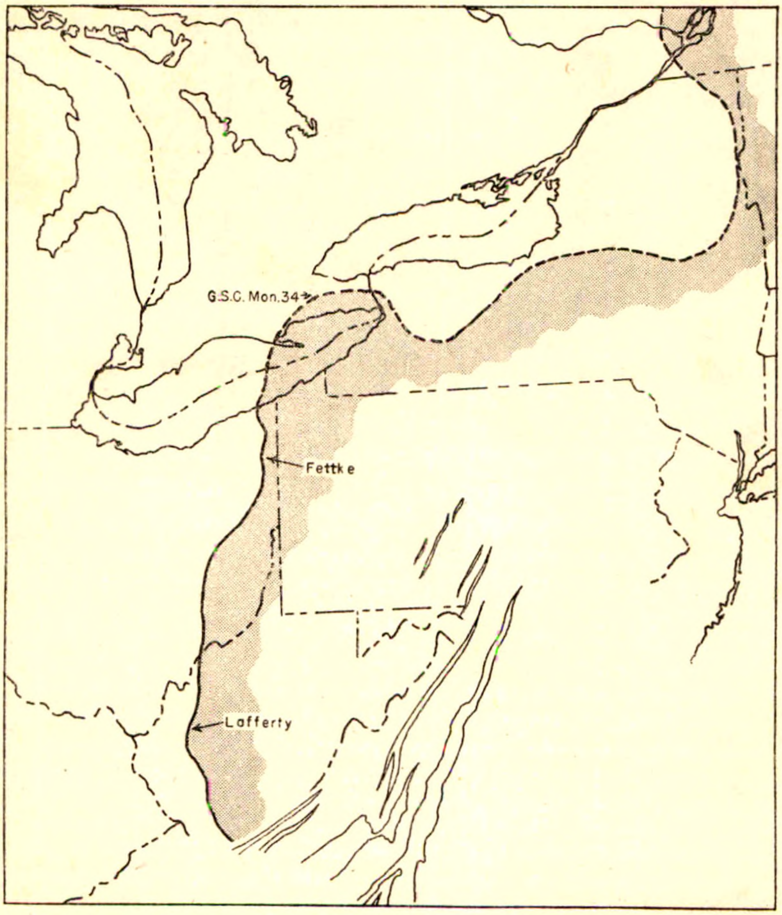

FIGURE 9.—*Margin of Oriskanian sediment in the Allegheny belt*

200 feet westward to the Cincinnati axis; thence Ulsterian carbonates increase to nearly 1000 feet in the Michigan basin, perhaps partially through overlap (Newcombe, 1933). Limestones continue southward into central Pennsylvania from New York without significant thickening; northeastwardly along the outcrop, they diminish across the "Harrisburg" axis or peninsula in east-central Pennsylvania (Willard, 1939) and exceed 500 feet in northeastern Pennsylvania, the lower beds being shale. Thus, in the Allegheny belt, the Ulsterian forms a plate of limestone, increasing in the western Champlain belt with addition of basal shales and diverging greatly into the Michigan basin.

Later Devonian series.—The post-Ulsterian sediments in the synclinorium are almost wholly terrigenous. The stratigraphy of the Erian in New York (Cooper, 1930) and succeeding Tully (Cooper and Williams, 1935) and Genesee (Grossman, 1941) have been described, as has that of the equivalent beds in Pennsylvania (Willard, 1939). There is general gradational overlap of the more sandy facies on the more shaly northwestward, with accompanying convergence (Reeves and Davies, 1937). The culmination is the spread of the continental facies in the late Devonian (Barrell, 1913-1914; Caster, 1934; Chadwick, 1933; 1936; Willard, 1939). Isopachal maps for Pennsylvania (Ashley, 1938) and West Virginia (Tucker, 1936, Pl. 1) emphasize the southeasterly tilting of the northern Allegheny synclinorium as the northwest quadrant of a great semicone of terrigenous materials. The source was highest and most eroded east of northern Pennsylvania.

SYNTHESIS OF EARLIER PALEOZOIC DEFORMATION

STRUCTURAL BELTS

The "major linear folds" of the present structure have been recognized and discussed by Keith (1923, p. 325); these have northeast trends as far west as the Cincinnati arch. In Ordovician and Silurian time, similar structural trend was dominant, but the alternate belts of relatively higher and lower structure have been considerably obscured by later deformation and burial. On the west, the Cincinnati axis, first structurally evident in the Silurian, was an arch only for a limited time and is on the margin of the Michigan basin. The Allegheny belt was essentially a structural plateau, the Rideauan trough within its eastern border being temporarily depressed. The Adirondack axis, first shown in the middle Ordovician, had Ordovician structure similar to that of the later Cincinnati axis; for part of Mohawkian time it was the site of an arch, but at other times the foreland margin of the deepening trough to the east. In late Ordovician (Richmond), it became domed relative to the surroundings in the region of the present Adirondack Mountains; normal faulting along the east flank and of the Laurentide dome to the northeast formed the present block-mountain structure, probably in early Silurian. Southwestward from New York, the Ordovician structural character has been obscured by later deformation and is not evidenced in younger sediments.

The Champlain belt to the east was a geosyncline until late Ordovician Cincinnatian and continued to subside more than its foreland. It was mostly destroyed and obscured by the Taconian orogeny, but the belt was overlapped from the northwest by later seas. Thrust sheets have buried the Quebec axis which is interpreted as having separated the

Champlain from the Magog trough to the east. Inasmuch as the Magog belt was thrust on the Champlain, their original relative positions can be represented only on a palinspastic map (Fig. 2). Thus, there were from west to east the following northeast-trending structural provinces or paleogeographic belts: Cincinnati, Allegheny, Adirondack, Champlain, Quebec, and Magog, differing in character and time of initiation.

GROSS DEFORMATION

The total deformation was downward in all parts of the Allegheny and contiguous belts through early and middle Paleozoic time, probably excepting the higher parts of the Silurian-formed Adirondack and Laurentide block mountains; that is, with minor exceptions, the surface on which Ordovician sediments were laid was far below the sea level of late Devonian time. The structurally higher areas only moved down less than the structurally lower. The principal sources of clastic sediments consistently lay to the southeast, being east of the Magog trough until at least late Chazyan time in the central Appalachians, until middle Trenton in the northeast. They were in the western Magog belt in the later Ordovician, within the thrust Taconian block in about the position of the autochthonous Quebec axis in Silurian, and in similar position but newly formed elevations in the later Devonian. The pre-Ordovician surface east of the Adirondack axis had been considerably deformed by late Devonian; much of that area, if not all, was above Devonian sea level.

REGIONAL TILTING

The Allegheny belt was tilted as a whole southeastward during times when uplift farther east mantled it with detrital sediment. Isopachs for each of the great clastic series are somewhat arcuate, the rocks forming the northern part of semiconical forms that center in the Champlain belt in eastern Pennsylvania. These reflect the presence of highlands to the east but also the influence of the Adirondack dome of the late Ordovician and the subsequent block mountains in inhibiting westward spread of sediment from New England. Depression was synchronous with deposition, and greatest where clastic sediments were thickest; deposition exceeded depression in latest Ordovician and late Devonian, prior to the Taconian and Acadian orogenies, causing shores to retreat northwestward from the advancing fluviatile plains. When terrigenous material was of limited amount, tilting tended to be southward, away from Laurentia, and marginward toward adjoining troughs.

Inasmuch as the greatest part of the thickness in the northern Allegheny synclinorium is in these Cincinnatian, Silurian, and Devonian detrital sediments, the northern part of the belt was tilted southeastward as the northern part of the semilenticular depressions.

THE AXES

The Adirondack and Cincinnati axes had arches for limited times (Pl. 3). The Adirondack arch is evidenced only during part of the Mohawkian epoch, and by domes in the late Cincinnatian, and the Cincinnati arch during the early Silurian Clinton, and possibly in late early Devonian. The deformation along the northern part of the Adirondack axis, in the Adirondack and Laurentide areas, formed block mountains in the early Silurian, and these stood persistently higher than the seas. The rock surface along the Cincinnati arch never stood markedly higher than on its flanks.

The axes margin an intervening Allegheny belt that tended to be less depressed than the Champlain trough and the Michigan basin except during orogenic times. Sediments tend to be thicker on the axes than in the intervening belt when adjoining troughs were greatly depressed, but the axes of present structures approach those of the early Paleozoic arches. In middle Trenton time, similar thickness of limestone was laid throughout the Allegheny belt along lines normal to it, but the beds on the Adirondack axis are terrigenous and thicker, diverging into the Champlain trough. Similarly, Cayugan carbonates are of fairly uniform thickness across the central part of the belt, but thicker on the Cincinnati axis, diverging into the Michigan basin. The Adirondack axis is one of oscillation in the Ordovician. Chazyan sediments are principally in the Champlain belt, Black River in the Rideauan trough of the Allegheny; the principal depression was on alternate sides. The Champlain belt contains much thicker sections of the middle Trenton; the gross result of these alternations is to have made the total thickness greater on the east flank than on the axis. Inasmuch as the Cincinnati axis separates the sinking Michigan basin from the southeast tilting along the margin of the semilenticular terrigenous masses of the Allegheny belt, there is alternation in maximum depression on the two sides between nonorogenic and orogenic times, resulting in less gross depression on the axis than on either flank.

The disturbances along the Cincinnati axis in Tennessee have been discussed by Wilson (1935). He concluded that, when "Appalachia" was relatively high, the Nashville dome was completely above sea level, for most of the time, or submerged only on the outer flanks, and that when "Appalachia" was low the dome was relatively low and was probably below sea level for most or all of the time. These relations might result from synorogenic upwarping of the dome or from eustatic withdrawal from an area of relatively higher surface; he attributed the sympathetic synchronism to the former. The sequence for the "Appalachian geosyncline" of Wilson is composite; some formations, as those of the Blount group, are in the Appalachian allochthone and were laid in the Champlain trough; others, as the "Trenton limestone," are nonterrigenous only in the Allegheny belt. McFarlan (1939) has questioned the presence of an arch in Kentucky in pre-Silurian and has shown that earlier structures were transverse (1938).

In Ontario, the first evidence of arching is subsequent to the Taconian revolution, though when highlands remained. Later history reveals some contrasting movements. The lower Lockport (Manistique) is limited to the Michigan basin and overlapped eastward by the Engadine, which forms a "plate." The Cayugan converges eastward across the axis, being thinnest in the middle of the Allegheny belt, evidencing depression of its margins. The late Cayugan probably offlaps each flank of the axis. The Helderbergian and Oriskanian are absent proximal to the axis, but the latter's margin in the Allegheny belt conforms to present structure (Fig. 9). Ulsterian limestone converges toward the axis from the west (Warthin, 1939) and probably overlaps. There being no recognized disturbance between the Taconian and mid-Devonian to the east, the secular movements cannot be called synorogenic, and the axis was more depressed than the belt to the east except in the early Devonian.

The basal Erian (Marcellus) is limy on the west, argillaceous to the east, and, though the contrast is not marked, later Erian continues limy in the Michigan basin. Inasmuch as in Senecan time terrigenous sediments invade the basin, the history is analogous to that along the Adirondack axis during the Mohawkian epoch, mudstones being restricted by the axis and subsequently passing beyond it. If the Erian influence on sedimentation is indicative of shallower seas over the axis, it may be called synorogenic, for it is synchronous with the advent of the Shickshockian disturbance, but there was not uplift, but lesser depression.

The two sides were synchronously depressed more than the axes only for limited times, and only the northern Adirondack axis had distinct uplift. Secular depression on alternate sides has been considerably responsible for the greater thicknesses on the flanks than on the axes of the present arches.

NORTHWEST TRENDS

Though the dominant trends in the region were northeastward, there were northwest-trending sags through the axes. The Montreal sag northeast of the present Adirondacks is suggested in the distribution of pre-Mohawkian rocks and antedated the formation of the Adirondack arch. Similarly, the Chatham sag between the present Findlay and Algonquin domes of the Cincinnati axis is suggested in thicker Mohawkian sediments, prior to the formation of the Cincinnati arch. The Frontenac axis at the outlet of Lake Ontario is first suggested in Oriskanian distribution. West of the Cincinnati axis, the Kankakee or Wabash axis has similar trend, separating the Michigan and Illinois basins; it had significance in the distribution of Mississippian sediments (Newcombe, 1933, p. 83). The principal structures of the southern Canadian Shield trend northwest and have been attributed to early Tertiary disturbance (Kay, 1942). Thus trends of some of the oldest and youngest structures cross those dominant in the Paleozoic. Keith (1923, p. 326-327) has discussed the cross folds, but to the writer their alignment is not so striking or significant as Keith believed.

TRANSVERSE MONOCLINES

The transverse monoclinal folds northeast of Lake Ontario (Kay, 1942, p. 632-636) have not been dated or explained. They resemble blocks tilted up to the northwest. The pattern seems en echelon, and suggests relative westward movement on the north.

EFFECTS OF DISTURBANCES IN ADJOINING REGIONS

INTRODUCTORY STATEMENT

Elevation-forming movements to the east of the Allegheny belt produced some or all of the following effects: (1) The uplifts (disturbances) were accompanied by erosion, yielding terrigenous clastic sediments that invaded the region. (2) Warping (epeirogenic) movements within and adjacent to the belt may have been associated with the disturbances. (3) The disturbances, and those in other parts of the earth, may have been accompanied by lowering of sea level (eustatic movements), the withdrawal or shallowing of seas influencing the distribution and facies of the rocks. (4) The loading of the area through any cause may have resulted in warping (epeirogenic) movements of isostatic adjustment. There were also epeirogenic deformations independent of the disturbances in eastern North America. The chronology is presented on the correlation table (Pl. 3).

TIMES OF DISTURBANCE

The character of highlands to the east can be judged by their erosion products. Uplift in middle Trenton (post-Shoreham) time filled the Champlain trough, which had previously received carbonates, with detrital materials; this was the Vermontian disturbance. The abrupt introduction of coarse clastics in eastern Pennsylvania and southeastern New York in late Maysville (post-Fairview) Ordovician indicates a second phase of uplift, the Oswegan disturbance; terrigenous rocks continued to spread to the west until the close of the period. The Taconian revolution (post-Richmond, pre-Alexandrian) was marked by folding and great overthrusting Whereas Ordovician sediment progressively coarsened, the Silurian is successively finer and less extensive; they form the "flysche" and "molasse" (Van der Gracht, 1931b, p. 998, 1011) of the Taconian revolution. From estimates of the quantity of material, the volume of rock eroded in the Ordovician was much greater than in the Silurian. The mass raised before the revolution was considerably greater than during the orogeny, particularly when it is considered that some of the Silurian clastics came from rocks raised above base level before the close of the Ordovician. That the elevation was greater preceding orogeny is not a necessary corollary, for there was recurrent uplift before and probably little thereafter.

The next great influx of terrigenous detritus was in the middle Devonian (post-Ulsterian); the later rocks generally coarsen upward and westward, the coarser and nonmarine facies gradationally overlapping the finer and marine. The sediments form the flysche from rising highlands to the east, whose uplift culminated in the Shickshockian (Acadian) disturbance; they are succeeded by Mississippian molasse.

South of the northern Allegheny synclinorium, there is evidence of an earlier Ordovician disturbance, for the Blount sediments of the Appalachian allochthone are thick and terrigenous; they are probably later Chazyan (Ulrich, 1929, p. 49) and came from highlands formed to the east in the Blountian disturbance.

In chronological order, the uplifts that influenced sedimentation in the Allegheny belt were the Blountian (late Chazyan), Vermontian (middle Trenton), Oswegan (post-Fairview Maysville) within the Ordovician, the Taconian at its close, and the Shickshockian (post-Ulsterian) culminating at the close of the Devonian.

CORRELATION OF MOVEMENTS IN THE ALLEGHENY BELT

Movements synchronous with disturbances have been postulated in regions proximal to those of principal displacement; "For these mainly epeirogenic, or moderately orogenic movements during orogenic periods, the term synorogenic has been used" (Stille, 1936a, p. 853). They may be reflected in new patterns of distribution and thickness, the formation or accentuation of troughs and arches.

There is notable change in distribution in the Allegheny belt from the Chazyan to the Mohawkian (Fig. 3d). In the former time, seas from the Champlain trough coursed into the northern part of the belt in the Ottawa Bay; but the Black River introduces the Rideauan trough, the axis of principal depression shifting from the Champlain trough to the east of the Adirondack axis; it is an example of oscillation. This change may have followed the Blountian disturbance, though the latter is not adequately dated; the Chazyan sediments in the northern Allegheny belt are Upper Chazyan (Valcour) and have been correlated with the Whitesburg and Athens formations of the Blount (Ulrich, 1929, p. 49). If this is correct, the epeirogenic movements in the north followed the introduction of Blountian terrigenous sediments in the southern Champlain trough; and the writer believes that there was similar oscillation along the Adirondack axis in the Appalachians.

Depression on both sides of the Adirondack axis in early Trenton time formed the Adirondack arch. This relative arching preceded the incursion of Vermontian clastic materials. That there was middle Trenton uplift, synchronous with stages of the disturbance, is shown in the truncation of older horizons by the Lacolle conglomerate northeast of the present Adirondack Mountains (Clark and McGerrigle, 1936) and relative uplift by the overlap of basal Denmark in the Mohawk Valley (Fig. 4). The axis of a later Trenton (Cobourg) arch was about 30 miles to the northwest. Movements along the Adirondack belt were not simple nor restricted to the time of Vermontian uplift.

Change in distribution of Cincinnatian sediments in the Adirondack region was synchronous with the Oswegan disturbance. The Lorraine diverges eastward regularly toward the axis and presumably across it; but the Richmond (Queenston) diminishes and disappears toward the southwest Adirondacks. The subsequent block faulting has been interpreted as resulting from "tensional stresses of the period following the Taconic [Taconian] revolution" (Megathlin, 1939, p. 119), though the dating is not conclusive.

The Cincinnati arch is first clearly evidenced in the early Silurian. Albion lithology changes from New York to Michigan, sandy facies being limited to the southeast of the Cincinnati axis; the thicknesses are similar on the axis and its flanks, and the relations are not demonstrative of a relative uplift of a linear belt or arch. In Clinton time, however, calcareous sediment in the Michigan basin was separated from argillaceous beds in the Allegheny belt along part of the axis; there was differential depression of the flanks. Though the first appearance of the arch is approximately synchronous with the Taconian revolution, it is not apparent until after the orogeny.

EFFECTS OF EUSTATIC MOVEMENTS

The demonstration of possible eustatic movements depends on the evidence throughout the earth.

> "If it can be shown that transgressions of the sea took place with simultaneous regularity in all the geosynclines of the world, except where modified by local causes, such as minor deformations, excess of sedimentation or other interferences, then the explanation for such processes can only be found in a rise of sea-level." (Grabau, 1938, p. 9)

The possibility that disturbances are accompanied by widespread marine withdrawals has been recognized.

> "If uplift of mountains in some geosynclinal belt is accompanied by crustal changes that for a time somewhat increase the capacity of ocean basins, it is easy to understand that lowered sea-level should result in simultaneous withdrawal of shallow seas from different continental platforms" (Moore, 1935, p. 1256).

There should be marine withdrawal at each known time of disturbance; excessive withdrawal and shallowing on relatively higher areas can be effects of uplift, and seeming lack of retreat and deepening effects of local sinking of greater amount than eustatic fall. Moreover, some withdrawals may be eustatic, but related to disturbances in other regions. Transgressions and regressions are shown on correlation tables (Pl. 3).

The times of Ordovician-Silurian disturbances can be rather closely analyzed and dated, but the phases of the Acadian disturbance of the Devonian are insufficiently known. Because of the proximity of the Allegheny belt to orogenic belts to the east, epeirogenic movements might be more prevalent than in the continental interior, but the proximity permits better correlation of disturbances with regional marine withdrawals.

The oldest Ordovician rocks lie in the northwest Allegheny belt and are lowest Canadian. The subsequent offlap, partial overlap, and offlap of later Canadian units is comparable to that in the interior (Stauffer, 1937, p. 61-62) and suggests eustatic control (Grabau, 1937, p. 735-741). There is no evidence of significant uplift in adjoining regions. The Chazyan sea retreated from this Ottawa Bay, and subsequently the Black River sea advanced. Though the withdrawal may have been synchronous with the Blountian disturbance, the latter seems to have preceded it a little, and there was significant Chazyan epeirogeny in the area. Eustatic fall possibly accompanied the disturbance, but evidence is inadequate.

The Trenton group lies disconformably on the Black River, the break being smallest in the Rideauan trough. Though this relationship is widespread and possibly eustatic in cause, no disturbance of this age is recognized in eastern North America. There was widespread retreat of the sea in middle Trenton (Denmark), somewhat after the Vermontian disturbance began. The younger Cobourg limestone overlaps the middle Trenton along the north margin of the Allegheny belt, and the equivalent Stewartville overlaps in northern Michigan; synchronous withdrawal is suggested in the Mississippi Valley (Kay, 1937, p. 294). The facies of the Denmark across the Adirondack axis show eastward deepening; hence there was independent epeirogeny. There is disconformity between the Cobourg and Upper Utica southwest of the Adirondacks, probably an effect of local uplift. Thus, the widespread withdrawal in later Denmark time is attributable to eustatic movement accompanying the Vermontian disturbance; movements along the Adirondack axis indicate local epeirogeny.

In northern Michigan and westward, Richmond sediments lie on the Trenton; to the east on Manitoulin Island, Eden and Maysville intervene. It is unknown whether the late Maysville overlaps or offlaps, and thus whether retreat was synchronous with the Oswegan disturbance. If there was eustatic fall in late Maysville, excessive rise followed, for Richmond beds overlie upper Trenton in the Mississippi Valley and northward to the Arctic; the seeming anomalous retreat westward in the Allegheny belt may be attributed to an excess of deposition over eustatic rise and epeirogenic depression.

There is persistent disconformity at the base of the Albion throughout the region, the oldest Silurian being unrepresented in the northern part of the synclinorium. If the Albion be the Castanea sandstone at the top of the Tuscarora in Pennsylvania (Swartz, 1934), the lower and main part of the Tuscarora is older than Silurian on the Cincinnati axis or in the northern margin of the synclinorium. Albion (Brassfield) limestone overlaps on the Cincinnati arch in Kentucky, Ohio, and Indiana (Foerste, 1935), evidencing a somewhat earlier presence of an arch. The pre-Albion (Cataract) disconformity could be due to eustatic movements accompanying the Taconian revolution, but the post-Albion arching is then not synchronous with the same revolution.

There was broad withdrawal in early Lockport, and probably at its close, but no disturbance is known in proximal regions. The later withdrawal of seas between Late Cayugan and Oriskanian might be related to the Caledonian disturbance of this approximate time in Europe.

In summary, there was withdrawal approximately synchronous with each disturbance. It is not known that each is universal, and hence that they are more than suggestively eustatic. The known evidence does not contradict the theory that eustatic fall accompanied each orogeny.

ISOSTATIC ADJUSTMENT

Consideration is next given to the possible effect of isostatic adjustment following loading. The position of deposition depends on other factors. The terrigenous sediments will be carried to areas that are low prior to sedimentation, from those that are high because they had been uplifted—original depression controls the place of deposition (Barrell, 1917, p. 748). Once an initially low area received sediments, the crust would be deformed downward if isostatic adjustment be immediate and local. However, the surface of deposition would rise, for, the mass for unit volume of sediments being less than that of the crust on the order of 5 to 6, there would be depression of the order of 85 feet for each 100 feet laid, and the top would stand some 15 feet above the original elevation. Any relationship in which sediments can be shown to have been laid at successively lower levels requires independent sinking, or eustatic rise. Further applied to the effect of rise of sea level, a change of 100 feet might result in the deposition of several hundred feet of marine sediments, with synchronous depression of the original surface of deposition (Hoffman, 1939, p. 1325). Thus the presence of 1000 feet of marine sediments would not require eustatic rise of more than some 200 feet.

The interpretation that an area has been differentially depressed by loading implies that more material has been laid in one place than another. The distribution of thicknesses and textures will be related generally to distance from and height of producing lands, and in the Allegheny belt clastics diverge toward highlands to the southeast, lines of equal thickness (isopachs) being arcs centered in eastern Pennsylvania. Gradational overlap of coarser and ultimately continental sediments over the marine in the later Ordovician and Devonian is in accord with the requisite rise of surface of deposition of the hypothesis. If loading alone were responsible, the thickness of several thousand feet in each case would require that the surface stood at an elevation of many hundred feet at the close of each period, disregarding eustatic changes, for the basal sediments in each terrigenous pile are of shallow-water origin. But the same result could be produced under any conditions wherein subsidence was slower than deposition, regardless of cause.

Sinking took place both in areas that received and lacked consequential terrigenous sediment. The Canadian and Chazyan carbonates of the Champlain trough approach a mile and lie between quite continuous units of Upper Cambrian and Mohawkian age that become contiguous to the west. The limy sediments of the Michigan basin converge greatly toward the margins and were not in proximity to highlands. The thickening of the higher Silurian (Tonoloway) limestone toward central Pennsylvania is similarly suggestive, though a part of the depression could be attributed to downwarping in areas contiguous to the Bloomsburg red shale delta, which in turn might have loaded the crust. Not only have these areas gone down, but those that received detritus have subsequently sunk. If late Ordovician rocks overlie a plane of deposition that sank because of loading, there was independent subsequent epeirogeny, for they are overlain by late Silurian and Devonian limestones. That this overlap was not due to eustatic change is shown in the relatively limited extent of equivalent deposits in other parts of the continent.

In nonorogenic times, the adjustments that had preceded are impressive. Despite great contrasts in amount of depression in the Allegheny belt during the late Ordovician, units like the Albion and Lockport are of remarkably uniform thickness and character over great areas. This balance might seem indicative of isostatic adjustment, but, if load controlled depression, surface of deposition should have risen toward areas of greatest thickness. On the other hand, if local sinking directed original deposition, the surface might be built to wave base of the textures represented, finer grades being carried to areas of lower surface of deposition. This should have led to a deficiency of mass in the areas of thickest sedimentation, except that the local sinking might have been induced by previously gained excess.

The conclusion from the record in the Allegheny synclinorium is that local sinking independent of terrigenous loading was a prominent factor in distribution of thicknesses, and that either local sinking complemented uplift, being greatest near the mouths of contributing streams, or that sinking resulted from isostatic adjustment to loading, with subsequent downwarping or erosion to a level in nonorogenic times. The close proximity of this area to the contributing highlands is not conducive to isolation of the factors. The best regions for analysis should be those in which deposits have been carried far from belts of uplift.

SUMMARY

The times of uplift east of the Allegheny belt can be determined from the times of incursion of terrigenous sediments. There were uplifts along axes and sinking of troughs at times of disturbance, but they were not limited to those times. There were marine withdrawals at the time of each disturbance, and these could have been eustatic. Evidence is not convincing that sinking of the crust beneath loaded areas was an effect of isostatic adjustment; and there was significant sinking that cannot be attributed to loading.

LATER PALEOZOIC HISTORY

Post-Devonian Paleozoic sediments lie south and west of the area of study, in Pennsylvania and Michigan; probably they were never deposited along the northern margin of the synclinorium. The lower Mississippian Kinderhookian and Osagian series are represented by the coarse clastic Pocono beds of Pennsylvania. These become finer westward, extending into the Michigan basin, which was separated by the

Kankakee arch from the Illinois basin to the southwest (Newcombe, 1932, p. 78-82). The fine clastic sediments grade into limestone along a belt passing through Illinois (Payne, 1940) and western Kentucky (Butts, 1917; Stockdale, 1939) to eastern Tennessee. The facies patterns are crescentic, with coarsest beds in eastern Pennsylvania nearest the highlands formed in New England in the Devonian. With reduction of the highlands, the earlier Mississippian molasse was succeeded by calcareous sediments. During most of the Meramecan epoch, the Cincinnati arch separated limestones and saline rocks in the Illinois and Michigan basins from limestones in a geosyncline with western margin passing through eastern West Virginia and extending southeastward beyond the limit of preserved rocks of this age. The later Meramecan limestones overlapped into southwestern Pennsylvania (Lucke, 1939) and westward over the Cincinnati arch (Stockdale, 1939).

The pattern of distribution of the later Mississippian and Pennsylvanian marks a significant change in trend from that of the earlier Paleozoic. This is evidenced in the Pennsylvanian Pottsville, which thins rather abruptly from more than 1000 feet to less than 400 feet in a belt passing from southern Pennsylvania through eastern West Virginia, eastern Kentucky, toward northwestern Alabama (Fig. 10). To the southeast, thicknesses exceed or approach a mile (Wanless, 1939; Butts, 1925; Prouty, 1923), whereas there is rather constant thickness of 200 to 400 feet through most of Pennsylvania, Ohio, northern Kentucky, Indiana, and Illinois. This belt of convergence, offlap, and overlap evidences a rapid sinking to the southeast. The structural contrast with its foreland can be compared to that of the middle Trenton between the Champlain trough and the Allegheny belt, though the latter has a much larger proportion of carbonate in the foreland facies. This late Paleozoic belt is designated the Virginian geosyncline; the trough seems to have been southwest-plunging. The western margin crosses that of the Champlain trough of the earlier Paleozoic in southern Pennsylvania and has suggestive relationship to the trend and character of structures formed in the Appalachian revolution. However, the Early Cambrian geosyncline also extended west of the Adirondack axis from Pennsylvania to Alabama.

Later Pennsylvanian and Permian rocks are preserved in diminishing areas in the West Virginia basin on the foreland of the Virginia geosyncline. There is insufficient preservation to permit reconstruction of the paleogeography. They become decreasingly marine and are wholly nonmarine above the Conemaugh (Reger, 1930), evidencing westward withdrawal of shores with the building of the terrigenous deposits.

CLASSIFICATION OF THE ALLEGHENY BELT

APPALACHIAN "GEOSYNCLINES"

The term geosyncline has been applied in the broadest sense to those regions of the earth that have undergone relatively great depression or sinking. Many authors include the Allegheny belt in the Appalachian geosyncline. The stratigraphy of the Paleozoic systems evidences the changing characters of the belts of downward movement and permits their classification.

The deformation of the Allegheny belt is separable into two principal classes of movements—those when terrigenous sediments failed to reach the belt and those when there were highlands to the east whose clastic detritus crossed the Adirondack axis. There was never significant terrigenous materials gained from the west and north; minor quantities are present in the base of overlapping sequences. When terrigenous rocks were absent or quite subordinate within the belt, the margining axes at times formed relatively low arches, as in early and earliest middle Trenton along the Adirondack axis and in Clinton time along the Cincinnati axis. When the troughs bordering the axes were sinking rapidly, the axes became flexures, and the Allegheny belt was relatively depressed along the margins, as in middle Trenton time along the Adirondack, and in Cayugan and Ulsterian times along the Cincinnati.

When there were highlands to the east, the belt was depressed in a semifunneled form, with the apex along its eastern margin, the greatest depression in the area near the terrigenous source and the deformation crossing the Adirondack axis with no appreciable influence. The deformation of the later Ordovician (Cincinnatian) (Fig. 6a) and later Devonian (Erian to Conewangan) (Fig. 10) is of this sort and affects the entire belt, whereas that of the Silurian involved only a part of the breadth. This deformation, attended by thick deposition, in the main has led to the reference of the belt to the Appalachian geosyncline. The belt contrasts with the geosynclines to the east and west, which have undergone great depression irrespective of the character of their sediments; as a whole, the Allegheny belt was never a trough.

The Michigan basin is an elliptical geosyncline (Newcombe, 1933, p. 101). During the Silurian to middle Devonian, it subsided considerably, though terrigenous sediments are practically absent (Pl. 2b). In the later Paleozoic, terrigenous sediments entered the area from the Allegheny belt across the Chatham sag in the Cincinnati arch; but independent subsidence continued during middle Mississippian (Meramecan) when detrital sediments were few.

The Michigan basin seems to have received the centrally thickening sediments because it was sinking, whereas the Allegheny belt was tilted to a considerable degree because of sinking synchronous with the production of terrigenous sediments to the east, beyond its margin. The former is a geosyncline defined by causes within its area; the latter was depressed in complement to uplift near by. The limits of the former are similar for a long time; the gross sinking of the latter is the sum of periodic movements. Both are within the craton or "shield."

The extracratonic Champlain trough was periodically sinking during earlier Ordovician, when terrigenous sediments were relatively unimportant, its western limit being sinuous. During later Ordovician, when thick terrigenous sediments were laid within, it continued to sink more rapidly than loading alone would cause if isostatic adjustment were immediate and complete. The eastern margin lies beneath the Taconic allochthone but can be interpreted from evidence in the easternmost autochthonous and westernmost allochthonous sections. It has been considered a submarine slope, a flexure with greater subsidence southeastward than in the Champlain belt (Bailey, Collet, and Field, 1928), or as a barrier separating two troughs (Ulrich and Schuchert, 1902, p. 639).

The boulder beds in the northwest marginal Magog trough sediments have been interpreted as submarine landslip deposits by those who believe the Quebec axis was a flexure (Bailey, Collet, and Field, 1928, p. 603-604). The writer concurs that in early Ordovician time there were no highlands along the axis, for, if there had been, there should be thick terrigenous sediments in the Champlain belt; but the presence (Raymond, 1913, p. 30) in local abundance (C. Faessler, personal communication, 1941) of crystalline boulders with exotic blocks of Lower Cambrian to Lower Ordovician limestone (Clark, 1924) such as would be found beneath the Taconic allochthone (Bailey, Collet, and Field, 1928, p. 601-602) require uplift along the axis with southeast sedimentary transfer of the rocks. They may have gained their sedimentary position by landslips but must have been raised before being transported and have been deposited originally in the margin of the Champlain trough. The Cow Head breccia of Newfoundland (Schuchert and Dunbar, 1934, p. 84-86) had similar paleogeographic position and has been attributed to "talus and landslides formed along a fault scarp that came into existence during mid-Ordovician orogeny . . . formed at the nose of thrust sheets"; it has been thought to be autochthonous. The post-lower Trenton Rysedorph conglomerate in the Magog belt southeast of Albany (Ruedemann, 1901; 1930) contains exotic boulders of Champlain trough facies (Kay, 1937, p. 276-277) and has been attributed to the "erosion of anticlinal ridges" (Ruedemann, 1930, p. 113).

The Quebec axis probably did not separate faunas, but facies, which in turn influenced faunas. There were uplifts along the axis, some of which may have risen above sea level, any of which must have directed currents and thus controlled distribution of facies. The conglomerates at Quebec were formed by movements antedating the Vermontian disturbance, for they are in Canadian beds and evidence the prolonged activity along the axis. The comparable "exotic" blocks in the northern Ouachita allochthone have been attributed to thrusting of rocks toward the geosynclinal margin (Hendricks, 1940); or landslipping from an archipelago, possibly bordered by normal faults (Kramer, 1933; Moore, 1934). The Haymond conglomerate of west Texas has boulders supposedly "derived from the erosion of thrust sheets advancing from the southeast" (King, 1937, p. 91), that is from the Ouachita geosyncline; their deposition antedates the important Marathon orogeny and thrusting. Those in the Ordovician Woods Hollow formation of the same region (King, 1937, p. 24-25) are of foreland facies. These are analogous to those along the Quebec axis.

A structurally active and relatively higher though perhaps interrupted barrier formed the eastern margin of the Champlain trough. To the west lay the Adirondack belt, alternately an arch and flexure. The trough sank endogenetically during Cambrian and earlier Ordovician time and became deeply depressed as it filled with terrigenous sediments from the Vermontian geanticline in the western Magog belt in middle Trenton.

CLASSIFICATION OF GEOSYNCLINES

Schuchert (1923, p. 195-197) has classed troughs such as the Champlain and Magog as monogeosynclines, formed by the partition of the earlier polygeosyncline, the "St. Lawrence Trough," by an intrageosynclinal geanticline, the Quebec "barrier." Stille (1936a, p. 517) has separated geosynclines into the greater intercontinental belts, orthogeosynclines, and those lying within the shield areas or cratons, "parageosynclines," the latter a term having a different earlier definition (Schuchert, 1923, p. 199). Inasmuch as the character of areas is transitory, any classification is fundamentally temporal. The writer adopts an Ordovician basis of reference for definition of paleogeographic belts; this may reflect his interest in the system, but it seems advantageous because the record is extensive, the belts had become distinct, and had not been altered by orogenies.

The Magog geosyncline has graptolitic slates, formed in at least moderate depth, and associated cherts and volcanic rocks, as in northern Newfoundland (Sampson, 1923; Heyl, 1936) and southeastern Quebec (Cooke, 1935; Clark and Fairbairn, 1936); they are in a belt later intruded (Keith, 1928). Lower Ordovician graptolites are of distinctive genera, and practically limited to the shaly and cherty sequences of the Magog belt, reflecting a structurally controlled facies. Similar rocks in the Ouachita area contrast with carbonates on the foreland, and they are found in the western belt of the "Cordilleran geosyncline" (Ferguson, 1924, p. 20; Ross, 1934, p. 942-945). That facies rather than land barriers limited their distribution is shown by thin graptolitic shales intercalated within the carbonates in Oklahoma (Decker, 1936), West Texas (King, 1937, p. 26-30), Utah (Clark, 1935), and southeastern Idaho (Mansfield, 1927); that local conditions changed, by the intercalation of thick graptolitic shales within limestone formations, as in British Columbia (Walker, 1926, p. 24-31; Evans, 1933).

The Magog belt lies in a pliomagmatic zone (Stille, 1936a), and the trough is a eugeosyncline.[3] Eugeosynclines are characterized by a prevalence of shales and cherts, the presence of volcanic rocks, ophiolites or greenstones (Stille, 1936a, p. 517; Bailey, 1936, p. 1717-1723); the rocks have been interpreted as formed at great depths (Ruedemann, 1936, p. 1545-1563) to an extreme that does not seem warranted by the coarser texture of some of the interbeds. They have been the locus of principal subsequent plutonic invasions.

In contrast to the Magog eugeosyncline, the Champlain belt contains dominant carbonates of shallow-water origin, unaffected by subsequent volcanism; it is a miogeosyncline. The lithologies are more like those on the craton than in the eugeosyncline, and with the craton, the belt is essentially amagmatic. But the miogeosyncline is extracratonic, inasmuch as the sediments in most regions have been thrust subsequently on the foreland.

Just as the Magog and Champlain troughs comprise the two parts of the St. Lawrence orthogeosyncline,[4] there are two similar stratigraphic belts in the Cordilleran orthogeosyncline. The western, eugeosynclinal belt has been classed as pliomagmatic (Stille, 1936b, p. 139-143) and called "Nevadan." Inasmuch as the name Nevadian has been applied to the intrusive disturbance at the close of the Jurassic period, it is confusing to use Nevadian or Nevadan for the stratigraphic belt. The lower Ordovician (Deepkill) graptolitic shales in Idaho (Umpleby, Westgate, and Ross, 1930; Ross, 1934) may be in the belt; equivalent rocks in southeastern Alaska (Buddington and Chapin, 1929, p. 72-79) are typically eugeosynclinal, and volcanic and cherty slates are prevalent in the late Paleozoic, Triassic, and Jurassic of western British Columbia. The eastern, miogeosynclinal belt has the well-described sequence of southeastern Idaho (Mansfield, 1927) and is the Cordilleran geosyncline of most authors. Keith (1928) emphasized the importance of subsequent plutonic intrusions in the western belt; Stille illustrates (1936b) the movements along the eastern border as westward, but the belt was thrust on the eastern belt (Umpleby, Westgate, and Ross, 1930, map; Merriam, 1940, Pl. 1), much as the Magog was thrust on the Champlain. The overthrust sediments of the eugeosynclinal belt form an unnamed allochthone inasmuch as the orogeny that produced the structure has not been dated; it is certainly post-Carboniferous and probably pre-Nevadian. Nolan (1929) has described the geosynclinal nature of the eastern belt in Nevada.

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[3] In reply to the writer's letters of November 8, 1939 and April 23, 1941, in which he requested a definition of orthogeosyncline and emphasized the contrasts between the stratigraphy of the pliomagmatic and amagmatic zones of Stille and the similarities of the stratigraphy of the latter with that of the craton, Stille, after discussion in a letter of March 20, 1940, proposed the following classification in a letter of May 24, 1941.

"1 Orthogeosynklinalen (alpinotype Geosynklinale) zerfallend in
a) eugeosynklinale Zonen (pliomagmatische Mutterzonen der Interniden)
b) miogeosynklinale Zonen (miomagmatische Mutterzonen der Externiden)
"2 Parageosynklinalen (germanotype Geosynklinalen auf schon konsolidiertem Untergrunde)."

The terms eugeosyncline and miogeosyncline have been defined recently (Stille, 1941, p. 15).

[4] An orthogeosyncline is defined on the character of the rocks and their history and is preferred to polygeosyncline, which is a geometrically descriptive term. Orthogeosynclines are polygeosynclines, but not all polygeosynclines are orthogeosynclines

The Allegheny belt seems comparable to the Coloradan, to the east of the Cordilleran belts, each having had at some times depressions filled with detrital materials gained from outside the belt; it is proposed that these intracratonic, terrigene-filled depressions be called deltageosynclines (Fig. 6; Pl. 2), in contrast to the endogenetic geosynclines like that in the Michigan basin that are termed autogeosynclines (Pl. 2b) in reference to their independent sinking and isolation from those near the more active tectonic zones.

In eastern North America, the sequence has been development of the orthogeosynclines, uplift of an intraeugeosynclinal geanticline Vermontia, folding and thrust of the eugeosynclinal Magog belt on the miogeosynclinal Champlain (Taconian revolution), intrusion within the eugeosynclinal (Acadian disturbance) and folding and thrust of the miogeosynclinal Champlain on its foreland, the Allegheny belt of the craton (Appalachian revolution). A comparable sequence seems represented in the Cordilleran belts in the Mesozoic and early Tertiary history. In the Ouachita region, magmatic invasions are not recognized, though the observable part of the belt is quite marginal; the thrusting is not dated with certainty, though it is at least in part intra-Pennsylvanian, and two belts of thrusting are not known. Exotic boulder conglomerates are present on the foreland margin of eugeosynclines—the source in the miogeosynclinal belts—and the deposits preceded the major thrusting. In a discussion of similar stratigraphy in the Ural region of eastern

Russia the writer (Kay, 1941b) misapplied the term orthogeosyncline to the more restricted eugeosynclinal belt.

The Ouachita and Virginian belts illustrate the confusion if one does not recognize the temporal nature of the belts. If one grant that the earlier Paleozoic rocks of the former are eugeosynclinal, though they lack characteristic volcanics, they are overlain by sediments of the Pushmataha series (Harlton, 1938, p. 853) that are like and presumably continuous with the lower Pennsylvanian sediments of the Virginian geosyncline (Van der Gracht, 1931) that is marginally intracratonic with reference to the earlier Paleozoic definition but is miogeosynclinal. On the other hand, the early Paleozoic "Ouachita" eugeosyncline must trend southeastward from Arkansas if it joins the extension of the Magog belt. It may be that the Pushmataha geosyncline is related to a different cause from the earlier trough and that the "Ouachita" geosyncline is composite. The great thrusting in the Virginian-Pushmataha belt may be limited to the area in which it lies on eugeosynclinal facies. That the Virginian miogeosyncline trended diagonally to the Ordovician Champlain trough has been stated (Fig. 10); whatever factors determined the trend and character of early Paleozoic belts had been superseded by others in late Paleozoic time.

Belts that had been sinking independent of terrigenous loading sank more when orogenic events produced quantities of detritus in neighboring belts. The Champlain trough developed this contrast in middle Trenton but retained its linear limits; and the later, transverse Virginian trough likewise maintained an essentially linear northwest boundary whether terrigenous rocks were unimportant or preponderant. These miogeosynclines contrast with the great semilenticular depressions of the simple deltageosyncline, or the more sinuous and varied outlines when the terrigenous sediments were distributed from a number of centers in other deltageosynclinal belts.

Foreland folds within the eastern border of the Allegheny belt failed to gain the magnitude of those of the comparable Coloradan belt of the late Paleozoic and particularly the Cretaceous; even during the late Gulfian epoch, uplifts within the Coloradan deltageosynclinal belt yielded significant quantities of clastic sediment (Knight, 1938); nor were they as important as the uplifts along the Wichita-Amarillo axis in the Ouachita foreland.

BARRIERS AND TROUGHS, AND OSCILLATION

Though the concept of barriers and troughs has been credited to Logan (1866, p. 45), the comprehensive application of the theory to the North American stratigraphy has been that of Ulrich and Schuchert (1902).

> "Though abundant corroborative evidence of the existence of a narrow barrier between the stratigraphically inharmonious areas is afforded by the structural geology, . . . the striking dissimilarity marking the faunas pertaining to the lithologically equally dissimilar Ordovicic rocks lying respectively on the east and west sides of the Great Valley emphasized the separate provinces" (p. 634).

The theory was elaborated by Ulrich (1911), who emphasized the "permanence of certain continental lines of elevation separating areas of depression" (p. 292) and introduced the principle of oscillation, "that the deposits on the opposite sides of the Appalachian barriers are as a rule not contemporaneous but successive" (1911, p. 292; 1929, p. 55). The barriers rather than the facies were considered directly responsible for the contrasting faunas (1911, p. 512-513);

> "correlations requiring the assumption of lateral changes in faunas caused by differences in character of bottom . . . are commonly found to be in error" (1916, p. 454). " . . owing to the overthrusting of the middle and eastern troughs in the Appalachian Valley, . . . the eastern edges of their respective formations are as a rule inaccessibly buried" (1911, p. 452).

The structures of the Appalachian revolution being transverse to the Ordovician, the relatively undisturbed area of the northern Allegheny synclinorium eastward reveals relations that may explain less evident relations in the thrust-separated sequences of the southern Appalachian Mountains. The character and influence of the Adirondack axis as a "barrier" has been studied, and the basis for the interpretation of the Quebec "barrier." The stratigraphic contrasts on the two sides of the Saltville-Helena thrust being comparable to those on the flanks of the Adirondack axis, and the latter being traceable toward the termination of the thrust, there are grounds for a southern "Rome Barrier" (Ulrich and Schuchert, 1902, p. 644) or "Cahaba Barrier" (Butts, 1927, p. 16), but only during Ordovician time; the principal contrasts across the thrust would be attributed to great displacement (Van der Gracht, 1931a, p. 111).

The writer interprets the "barriers" as having influenced facies through controlling distribution of sediment and attributes many of the faunal contrasts to the facies and original geographic separation. Thus, Ulrich considered that the Athens shale was laid in the "Athens Trough," continuous with the "Levis Trough" in which the homotaxial Normanskill was laid; but the former was in the southern continuation of the Champlain trough, and the latter in the Magog to the east, the Blountian disturbance having led to the deposition of sediments in the south comparable to those in the Magog in the north. Five troughs were postulated between the Adirondack and Green mountains of New York-Vermont (Ulrich, 1911, p. 442-446); the rocks of the third are those of the Taconic klippe, those of the first, second, and fourth are continuous in field outcrop (Cady, 1938), and the fifth includes metamorphic rocks considered (Keith, 1932, p. 404-406) to be of the same sequence as the third. As now understood, the first, second, and fourth are parts of the continuous section of the Champlain trough, and the third and fifth were laid in the Magog.

The idea of multiplicity of troughs seems to have been developed from two independent considerations. First, having concluded that some thrusts are along barriers, it was assumed that most or all are along barriers, and that each thrust slice corresponded to an original trough. Second, in the belief that the barriers persisted through the Paleozoic, each barrier might be based on evidence from a single time and interpolated into the history of all periods. It is now thought that the later troughs had independent limits and more westerly trend than those of the middle Ordovician. The "transverse axes" seemed evidenced by related phenomena; inasmuch as the later troughs cross the earlier, Lower Devonian sediments, for example, are present in the Champlain belt in the north but absent in the south because the Devonian trough crossed the Adirondack axis, by then dormant, into the Allegheny belt. The disappearance of units is further related to thrusting. Just as the rocks of the Champlain belt pass beneath the Taconic allochthone in Quebec (Bailey, Collet, and Field, 1928, p. 602-604) and New York, so also those of the Champlain belt in the western slices in northern Virginia pass beneath more easterly thrust sheets south of Roanoke, and the eastern Allegheny belt sediments beneath the Appalachian allochthone in the southern Appalachians. Moreover, the Champlain belt sediments may have been eroded off the more easterly thrust sheets southward inasmuch as the belt passes from the early Devonian trough in the north to the region east of the trough in the south.

Ulrich's (1911, p. 415-416) evidence for oscillation for the Adirondack area is principally correct (Fig. 4), and there seems comparable foundation for the Chambersburg example (p. 321-329). These may be more exceptional than he believed, however. Some of the evidence that seemed significant is related to the changing trends of later troughs.

In summary, the writer believes that there were a few barriers and troughs in the Ordovician and that faunas were controlled and limited by facies, which in turn were dependent on fundamental structural changes. The middle Ordovician troughs did not persist long, but others, of more southwesterly trend, followed. The Appalachian structures seem directed by their presence or by the more fundamental causes of their origins. Oscillation is represented in the stratigraphy but is not thought to have been the principal cause of contrasts. Appalachian thrusts have a major part in the distributions now apparent, just as have the earlier Taconian thrusts.

APPALACHIAN REVOLUTION

THE FOLDS

The Appalachian revolution came at or near the close of the Paleozoic era. The more obvious and spectacular effects are in the Appalachian Mountains from eastern Pennsylvania to Alabama, but folds of some magnitude extend into the Allegheny synclinorium of northwestern Pennsylvania and southern New York. The trends are northeast in western Pennsylvania (Sherill, 1934; Cathcart, 1934; Fettke, 1938, p. 152-154), becoming nearly east in southern New York (Wedel, 1931; Fox, 1932). The northwest limit of folds having structure of more than 100 feet passes through southern Cattaraugus, Allegany, and Livingston counties, New York; to the north, the regional structure is essentially homoclinal. The asymmetry of northern folds contrasts with that in the more strongly folded Appalachians in that the steeper flanks face southeastward rather than northwestward; the significance has been discussed (Sherrill, 1934). The trend of Appalachian folds in New York departs by nearly 45° from that of the Adirondack axis, the divergence decreasing southward.

The northern Appalachian foreland folds parallel the Appalachian Structural Front (Price, 1931). The front lies west of the Virginian geosynclinal margin in the south, and far to the northwest in Pennsylvania (Fig. 10). It does not follow any recognized axis, flexure, or trough of the Paleozoic, and the origin of its specific position is not evident. It does lie in the position of thick Paleozoic sediments, but these vary in manner of deposition, the principal thickness in the north being in the Ordovician and Devonian deltageosynclines and in the south in the Virginian miogeosyncline. The generalization is not definitive of the line of contrast at the front.

THRUST FAULTS

The folded and faulted western Appalachians lie in the Allegheny belt; faults of considerable horizontal displacement pass into folds at their ends. The Saltville and Helena faults separate middle Ordovician rocks comparable to those in the Champlain and Allegheny belts farther north; and the northern termination of the Saltville thrust is about on the extension of the Adirondack axis. Though stratigraphic contrasts are small across the Saltville thrust as it approaches its termination in Virginia, they are great across the Helena thrust in Alabama (Butts, 1927, p. 16); the contrast in Alabama has been attributed to the relatively great displacement (Van der Gracht, 1931a, p. 111-112). The Saltville-Helena thrusts form the margin of the Appalachian allochthone, in the sense that they separate two stratigraphic provinces, of which the margin of the eastern has been thrust on that of the western. The allochthonous rocks of the Saltville slice pass northeastward into the Champlain belt where they are autochthonous with respect to the Taconian thrusts, and rocks eastward to the Martic thrust were laid in the Champlain belt. Southeasterly fault lines extend progressively farther northeastward in Virginia (Butts, 1933, map), comparably to the southward termination of the principal thrusts in the Champlain belt of western Vermont (Cady, 1938).

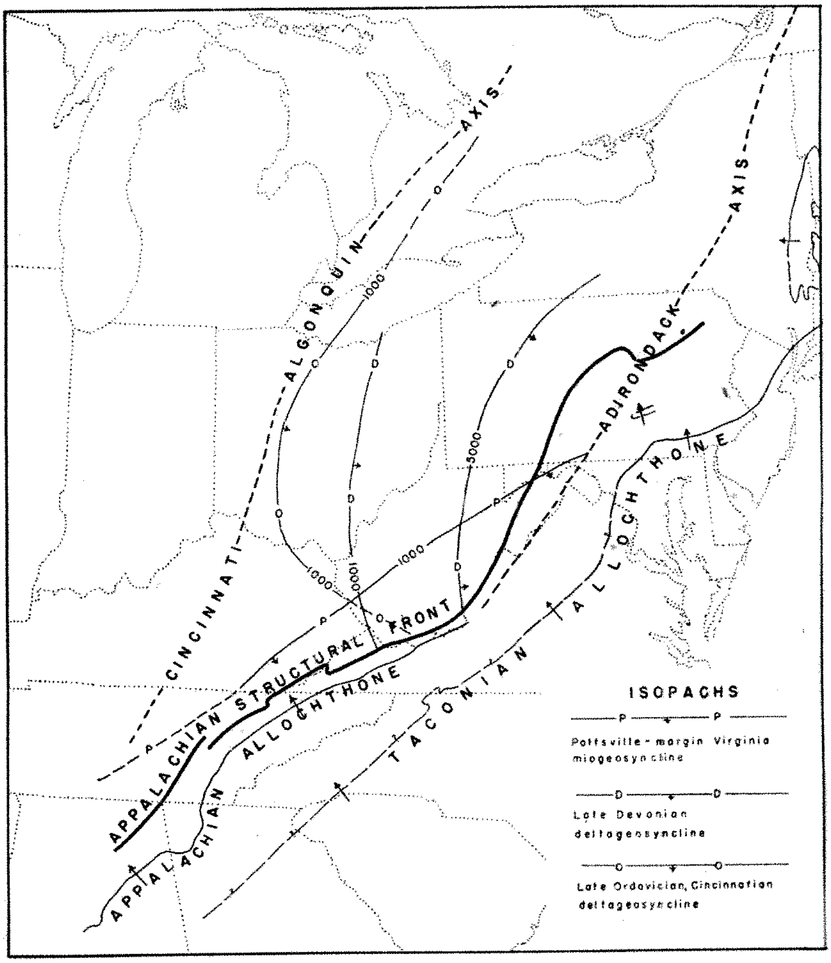
ALGONQUIN AXIS
CINCINNATI
ADIRONDACK AXIS
APPALACHIAN STRUCTURAL FRONT
APPALACHIAN ALLOCHTHONE
TACONIAN ALLOCHTHONE
1000
5000
ISOPACHS
P P
Pottsville - margin Virginia miogeosyncline
D D
Late Devonian deltageosyncline
O O
Late Ordovician, Cincinnatian deltageosyncline

FIGURE 10.—*Appalachian structural front with reference to Paleozoic geosynclines*

The western fault lines of the Appalachian system are similar to the margin of the Virginian geosyncline in position and trend, but those farther southeast increase in throw southwestward, and thus the rocks were originally farther east in the south; the trend of the original breaks was more northerly than the present fault lines. Relatively great displacement toward the south has rotated the slices so that the fault lines almost parallel the margin of the Virginian belt, but the trend of the initial fractures must have been more southerly. The thrust along the margin of the Appalachian allochthone developed along the Adirondack axis. This axis parallels the Cincinnati axis in the north, the two lying 300 miles apart from New York to Virginia. If the parallelism continued southward, the displacement on and west of the Helena thrust in Alabama would be 100 miles or more. The amount seems unreasonably large, the two axes were structurally active at different times, and their parallelism may be coincidental.

Keith has stated that

> "identity of the belt of overthrusts with the belt underlain by thick sections of weak Cambrian or Algonkian beds and floored by a . . . slope of the pre-Cambrian granite and gneiss is so complete that they must be acknowledged as cause and effect" (1928, p. 346).

Considering the continent as a whole, this generalization is warranted, in spite of the fact that the extent and western margin of the earliest Paleozoic Cambrian "geosyncline" in the Appalachian region is unknown but is far to the west of the zone of the principal thrusts; nor is the extent of the pre-Cambrian geosyncline known. The pre-Paleozoic rocks in the Taconic and Martic thrust sheets are metasediments where exposed, but unfortunately the limits are poorly defined westward. There are known areas of granitoid rocks in the Champlain belt in the north; the Taconian thrust sheets developed in the eugeosyncline, and possibly the pre-Cambrian metasediments were laid in the same belt; the longitudinal extent of the Glenarm series of metasediments and their essential parallelism to the Taconian thrust lines are certainly suggestive.

Ashley has written that "apparently the great thrust moved somewhat crosswise to the original lines of sedimentation, which extended more nearly north and south, as suggested by the [Devonian] isopachous lines" (1938, p. 426). Inasmuch as the latter have arcuate pattern when considered as a whole and reflect semilenticular depression, the divergence in trend varies with points of reference; but the isopachs and the Appalachian Structural Front in Pennsylvania strikingly approach parallelism (Fig. 10). The front roughly parallels the margin of the Virginian belt in the south and the deltageosynclinal isopachs in the north.

The line of the eugeosynclinal margin corresponds to that of initiation of the Taconian thrusts whatever the reason for the specific position;

and the margin of the Champlain belt corresponds to that of the thrusts on the margin of the Appalachian allochthone. The cause would seem more fundamental than the character of the surficial rocks; their nature may reflect the same basic structural conditions. Similarly, the principal Laramide thrusts formed within the eastern margin of the Cordilleran miogeosyncline (Keith, 1928, p. 333-334), and others, perhaps older, along its western margin (Merriam, 1940), east of the eugeosyncline.

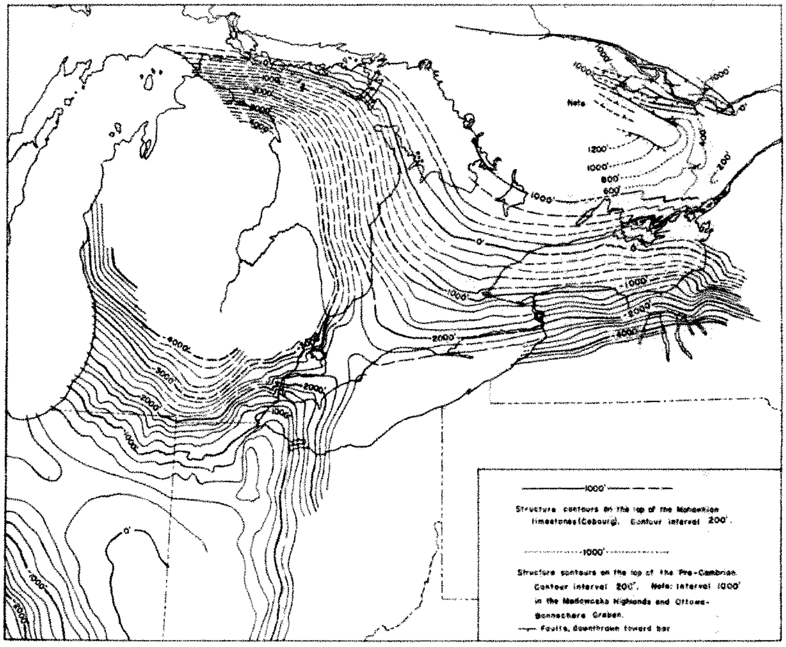
1000'
Structure contours on the top of the Mohawkian limestones (Cobourg). Contour interval 200'.
1000'
Structure contours on the top of the Pre-Cambrian. Contour interval 200'. Note: interval 1000' in the Madawaska Highlands and Ottawa-Bonnechere Graben.
Faults, downthrown toward bar
Note
1200'
1000'
800'
600'
0'
1000'
2000'
3000'
4000'
5000'
6000'
200'

FIGURE 11.—*Structure contours on top of Mohawkian limestone*

LATER DEVELOPMENTS OF THE NORTHERN ALLEGHENY SYNCLINORIUM

The minor transverse monoclinal folds within the northern margin of the synclinorium (Kay, 1942, p. 632) are of unknown age and origin. The border is essentially homoclinal north to the uplifted fault block on the south side of the Madawaska Highlands of Ontario. The region to the north has been extensively block-faulted by movements attributed to the close of the Mesozoic era or the early Tertiary. Subsequent denudation implies Tertiary uplift of the whole region. The present state is shown in a structure contour map on the top of the Mohawkian limestones (Fig. 11).

SUMMARY

The Allegheny synclinorium is a structural belt of complex origin. It was first defined on the east by the formation of the Adirondack axis in the Ordovician and on the west by the initiation of the Cincinnati axis in early Silurian. The great thickness of sediment is principally an effect of its position in the northern quadrant of great semilenses of terrigenous rocks laid in Ordovician, Silurian, and Devonian deltageosynclines; otherwise it stood relatively higher than the Champlain trough and the Michigan basin on the two sides. The folds of the Appalachian revolution invade it in the south, and the southern synclinorium is partially buried beneath the Appalachian allochthone, the bordering fault having originated along the original southeastern margin of the Allegheny belt. Though the Frontenac axis in the northeast seems of earlier Devonian origin, there was significant block faulting along the northern margin in late geologic time.

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WEST TEXAS BARRED BASIN

BY ROBERT I. ROTH

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ABSTRACT

The regional stratigraphy of the Leonard series along the eastern portion of the Permian Basin of West Texas is set forth by means of a number of strategically placed stratigraphic sections. Important faunal zones are indicated and discussed. Many characteristics of a barred basin environment are reviewed, and the hypothesis is advanced that the Leonard was deposited in a barred basin.

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The writer began surface and subsurface studies of the Permian Basin in 1929 while working for the Indian Territory Illuminating Oil Company of Bartlesville, Oklahoma. This company gave every possible aid, which is gratefully acknowledged. Mr. John W. Skinner has identified or checked the identification of all fusulinids found. Without this help regional correlations would be difficult. Drs. A. K. Miller, R. C. Moore, and P. B. King have read the manuscript and offered many valuable criticisms. Finally, all help necessary to the further study of the problem and permission to publish were given by the Humble Oil & Refining Company of Houston, Texas.

INTRODUCTION

The peculiar sedimentary record preserved in the Permian Basin of West Texas has elicited many explanations both for the environment and the source of these sediments.

In the Glass Mountains, the Leonard limestones, sandstones, conglomerates, black shales, and bioherm masses represent a normal though shallow marine environment. To the northeast and north they give way to black shales. Continuing along the above directions these shales grade rapidly into beds of limestone, dolomite, anhydrite, sodium chloride, reddish brown to greenish gray shales, and grits. Certain horizons show minor sandstones and conglomerates. The composite sedimentary aspect compares favorably with those of barred basins and indicates a similar environment.

ZONE OF *PERRINITES* OR TYPE LEONARD

The zone of the ammonoid *Perrinites* is defined by Miller (1938, p. 1014-1019) to comprise beds belonging to the Middle Permian Leonard series (Fig 1). Sediments of the zone represent almost every type of environment and facies—normal marine, euxinic, and desiccated—and no known barrier separates the several types.

The Leonard, recently elevated to the rank of series, is one of the stratigraphic units of West Texas which were defined and named long before their faunal content was clearly understood. At Leonard Mountain, its type locality, it is bounded by an unconformity at the base and a disconformity at the top. Faunally it is characterized by *Perrinites* and *Dictyoclostus bassi* McKee.

Fourteen miles northeast of Leonard Mountain a wedge approximately 550 to 600 feet thick is included at the base of the Leonard because there is no obvious break at its top. This wedge, not present in the type area, is characterized by *Schwagerina crassitectoria* Dunbar and Skinner.

	TEXAS				
	AMMONOID ZONES	GUADALUPE AND DELAWARE MOUNTAINS	GLASS MOUNTAINS	NORTH CENTRAL TEXAS	FUSULINID ZONES
PERMIAN	CYCLOLOBUS	XXX	XXX	XXX	
	TIMORITES	CAPITAN UPPER DELAWARE	CAPITAN	CUSTER IN PART	POLYDIEXODINA
	WAAGENOCERAS	MIDDLE DELAWARE LOWER DELAWARE	WORD	XXX	PARAFUSULINA
	PERRINITES (BY DEFINITION)	BONE SPRING	LEONARD	PEASE RIVER CLEAR FORK	
		X	XX	WICHITA	SCHWAGERINA *CRASSITECTORIA*
	PROPERRINITES	WOLFCAMP X	WOLFCAMP	WOLFCAMP	PSEUDOSCHWAGERINA
PENN.	UDDENITES	MAGDALINA X	GAPTANK	CISCO	TRITICITES
	X NOT EXPOSED XX PRESENT XXX ABSENT				

FIGURE 1.—*Permian ammonoid-fusulinid zones (Texas)*

Neither *Perrinites, Properrinites,* nor *Dictyoclostus bassi* McKee have been reported from it. In boring No. 19 in Irion County, 175 miles northeast of Leonard Mountain (Pl. 1), *S. crassitectoria* has been found to range through an interval of about 600 feet. In boring No 18 in Tom Green County, 23 miles north and a little east of No. 19, the same interval is 620 feet. Apparently the *S. crassitectoria* zone maintains a remarkably uniform thickness over great areas. *S. crassitectoria* has not been reported from any well north and east of boring No. 18.

On the east side of the Permian Basin about 600 feet of strata lying above the Wolfcamp, as now defined, contains the ammonoid *Proper-*

rinites (Fig. 1). These strata, which are called the Wichita group, may be traced lithologically from well-to-well southwestward into the zone of *S. crassitectoria*. Therefore, the lower part of the Leonard series, as defined in some areas, but seemingly not in the type section, may be older than the zone of *Perrinites* and may contain *Properrinites*. If this should prove to be the case, the zone of *Perrinites* is not synonymous with the Leonard series as this term is now being used. Proof of the supposition just offered is lacking, however, and the author follows Miller's definition of the zone of *Perrinites*.

ECONOMIC ASPECTS OF THE LEONARD

The importance of the economic aspects of the Leonard cannot be over-emphasized for a large percentage of its commercial oil and gas seems to have originated therein. Immediately preceding the deposition of the Leonard a broad emergence developed continuous porosity by weathering and alteration of the underlying limestones into dolomites and porous limestones that now serve as reservoirs Very little oil or gas has been discovered in that portion of the Leonard underlying the San Angelo horizon. On the east side of the Permian Basin proper a few small areas have yielded oil in commercial quantities from strata above the base of the San Angelo. One of these producing areas is in Garza County. Over most of the Permian Basin, pre-Custer weathering developed continuous porosity and dolomitization in the limestones beneath the Custer. These magnesian limestones, some of which are younger than Leonard, have been loosely referred to as the West Texas big lime. Locally porosity has developed through a stratigraphic interval of 300 feet or more and in many places this porosity is filled with oil. Very little oil or gas has been found in the areas where the big lime grades into strata which are predominantly chemical deposits such as sodium chloride, anhydrite, and primary dolomite.

MAJOR SUBDIVISIONS

GENERAL STATEMENT

On the east side of the Permian Basin, the Leonard may be subdivided into the Wichita, Clear Fork, and Pease River groups (Pl. 1). Their characteristics, both on the surface and in the subsurface, are as follows, starting with the oldest.

WICHITA GROUP

The beds from the top of the Elm Creek limestone upward to the top of the Lueders limestone are assigned to the Wichita group. A regional unconformity marks the base of the group, and probably there is a disconformity at the top. The average thickness is 675 feet. In the type area, the group consists of red clastics which are very difficult to separate from redbeds of the overlying Clear Fork and underlying Cisco. The topography of the Wichita redbeds is rolling and broken by many local badlands; that of the marine Wichita beds is rough on account of strong development of stairlike escarpments.

In the subsurface, the Wichita grades from red clastics into evaporites and marine sediments. At or near the base is a widespread zone of anhydrite. Where the Clear Fork becomes marine, the top of the underlying Wichita is difficult to recognize for in these areas its deposits are also marine.

CLEAR FORK GROUP

This group includes all beds from the top of the Lueders limestone up to the base of the San Angelo sandstone. Disconformities probably occur both at the top and bottom. The average thickness is 1150 feet. Except for a few magnesian limestones, which are exposed in the southern area, practically the entire section consists of soft reddish-brown clastics. A few beds of gypsum and anhydrite are exposed in the upper portion. The Clear Fork outcrop is a monotonous plain interrupted locally by badlands.

In the subsurface the Clear Fork red clastics grade westward into marine sediments and evaporites. Where the San Angelo changes to dolomites and marine sediments, the top of the Clear Fork is difficult to recognize because in these areas it also is marine.

PEASE RIVER GROUP

The rocks of Leonard age that overlie the Clear Fork group unfortunately have been designated as the San Andres group, the El Reno group, the Pease River group, and the "Blaine of Texas". Each name has advantages and proponents. The Pease River group is adopted here because its upper and lower limits are easily recognized where it crops out on the east side of the Permian Basin. The same cannot be said of the other groups when traced into the area. The Pease River group includes all beds from the base of the San Angelo sandstone up to the base of the Custer which is 10 to 30 feet below the Childress magnesian limestone. The group is disconformable at its base, and a regional unconformity occurs at the top. Average thickness is 870 feet. Surface exposures show many cycles of evaporites, which are now greatly reduced by solution. Except for the area in Nolan County, where red clastics are relatively prominent, the outcrop is uniformly rough with a pro-

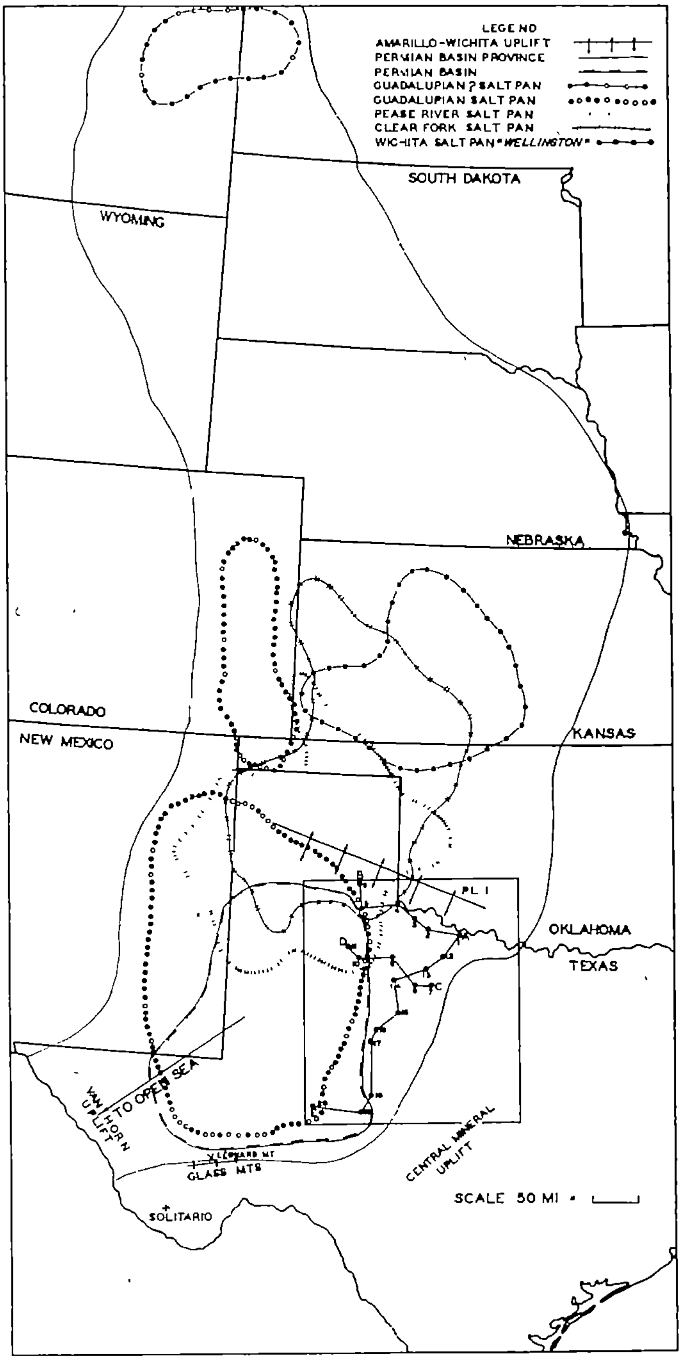

FIGURE 2—*Permian salt pans*

nounced development of stairlike escarpments, and vegetation is notably different from that of adjacent groups.

In the subsurface clastics gradually diminish basinward, and the recurrence of evaporites is clearly marked, inasmuch as solution has neither removed the sodium chloride nor hydrated the anhydrite.

Figure 2 generalizes the location of the various salt pans in the Permian Basin province. The salt pans are not necessarily massive salt beds but may only represent salty shales. No attempt was made to indicate all the pans found in the Guadalupian.

PALEOGEOGRAPHIC SETTING

The midcontinent as a whole has certain geological characteristics that set it apart from the rest of the United States. Crustal stability has dominated since the beginning of the Cisco. Orogenic movements broke this stability at but one place—the Amarillo-Wichita uplift—which was slightly rejuvenated during post-Permian time. Clastics were derived from the ancient Amarillo-Wichita mountains and form a large percentage of the Wolfcamp and Cisco series. All traces of this positive area were obliterated by burial under the Leonard (Fig. 2).

During Leonard time the Permian Basin province seems to have been a vast peneplain bordered by an upland of slight relief. Except for the widespread San Angelo sandy deposits, clastics coarser than shale are found only on the margins of the Permian Basin. These clastics attain considerable thickness in four main areas: the Anadarko Basin of west-central Oklahoma; Coke and Nolan counties, Texas; Marathon region of West Texas; and the northwestern counties of the Texas Panhandle. The latter area extends westward into New Mexico and probably is related to the Middle Permian Coconino and DeChelly sandstone deposition in Arizona.

Extremely uniform deposition during Leonard time is indicated by lithologic units which range from less than a foot to over 20 feet in thickness and which have been traced for hundreds of miles. Thickening or thinning is of minor importance throughout the area. Seemingly any positive or negative regional movements were epeirogenic, and accordingly small negative movements would cause regional inundation.

All paleontologic and stratigraphic evidence indicates that connection of the Permian Basin with the open sea was toward the south and southwest.

The Llanoria geosyncline, which was filled with thick Ordovician to early Pennsylvanian sediments, extends into Texas from southeastern

Oklahoma. Sediments preserved in this geosyncline have been traced around the east and southeast portion of the Central Mineral uplift of Texas, and thence westward into the Marathon region (Sellards, 1932). Their last appearance is in the Solitario uplift, located in the extreme southwest part of West Texas. A peculiar lithologic facies aids in identification. Black to greenish-black shales associated with coarse and fine sandstones, which in most places are rust brown to greenish gray, overlie a thick succession of cherty limestone. All are now complexly folded and somewhat metamorphosed. Cherty and novaculitic limestone beds, more than 1000 feet thick, dip at high angles and profoundly affect the topography. They provide much of the Permian conglomerates of the Marathon region. The southwesterly open sea connection of the Permian seas must have crossed the sediments of the geosyncline, and because of their geographic position the cherts and novaculites offered effective barriers to an open sea connection, thus affording an ideal area for the growth of bioherms. Permian sediments in this and adjacent areas exhibit a great variety of animal and plant growths whose development would tend greatly to restrict the existing channels to the open sea. The bioherms would also tend to maintain themselves if a negative movement developed.

ENVIRONMENT OF DEPOSITION

The midcontinent apparently was a vast desiccating barred basin during the deposition of the Leonard and, in fact, during most of the Permian. In the southern portion of the Permian Basin there is an area of black shale deposition in which the entire Permian section from Wolfcamp through lower Guadalupian consists of bituminous shales. It is therefore very probable that this area of black shale deposition did not rise above the level of the seas during the time indicated. A few thin brownish-black cherty limestone are scattered through the section. Pyrite is present in many zones. The few sandstones present are very fine, bordering on a grit. Light gray limestones surround the black shale on the east, north, and northwest. The transition from the euxinic environment to one of desiccation is not extremely sharp, for it extends through nearly 60 miles. Continuing in the three directions indicated, dolomites replace the limestones, and farther out these in turn are broken up by interfingering greenish-gray shales and anhydrites. The latter transition is best developed to the north. In the peripheral zone, reddish-brown shales and evaporites dominate the whole section. Thus, outside the "black shale" basin the normal sequence of deposition is gradational, both vertically and horizontally (Pl. 2).

An ideal cycle of chemical deposition of evaporites is as follows:

Sodium chloride (does not crop out because of solution)
Anhydrite (usually hydrated on the outcrop)
Dolomite or magnesian limestone (locally fossiliferous)
Apple-green shale
Blue-green shale
Reddish-brown shale

The thickness of each member of the cycle may vary greatly, but each is usually present. Sand is common in the shale but rare in anhydrite. Cycles are best developed in the Pease River group.

Several hypotheses have been advanced to explain the environment of Permian Basin deposition. One pictures a desiccating sea slowly retreating southwestward. This concept has several drawbacks (Fig. 2). The Leonard seems to have had a wider distribution in the Permian Basin province than any other series except possibly the Guadalupian. Little evidence is found to indicate that the Pease River group was more restricted than either the Clear Fork or Wichita groups. There is a marked tendency for the more soluble evaporites to be confined to an area north of the "black shale" basin. The position of the latter directly in the path of an open sea connection complicates the hypothesis of a southwestwardly retreating sea.

Deposition of black shales of the type found in the Permian Basin requires very specialized conditions that seem to be closely related to the growth of disconnected limestone masses. Mention of bioherms is common in the literature referring to the Permian Basin. The existence of bioherms connotes the presence of lagoons. Lagoonal deposition has been studied from many angles, but explanation is lacking for the many types of deposition found to date. Woolnough (1937) and Twenhofel (1939) have described conditions of deposition remarkably similar to those here included in the Leonard and have advanced the descriptive term barred basin for such deposits.

SECTION A-B

LOCATION OF BORINGS

No. 1.—Humble Oil & Refining Company's Williams & Parker No. 1, Wichita County, Texas. 200 feet from South and East lines of west 213.3 acres, section 25, H. & G. N. R. R. Survey. Elevation, 1029 feet.

No. 2.—Humble Oil & Refining Company's B. D. Barker No. 1, Wilbarger County, Texas. 1650 feet from South and 330 feet from West lines of SE¼ of section 31, block 9, H. & T. C. Survey. Elevation, 1250 feet.

No. 3.—Amerada Petroleum Corporation's J. S. Rice *et al.* No. 1, Hardeman County, Texas. 1635 feet from North and 1679 feet from West lines of section 81, block H., W. & N. W. Survey. Elevation, 1476 feet.

No. 4.—Hines *et al.* W. T. Walling No. 1, Hardeman County, Texas. 1900 feet from the South and 2100 feet from the West lines of section 368, block H., W. & N. W. Survey. Elevation, 1845 feet.

No. 5.—Phillips Petroleum Company's A. C. Hughes No. 1, Hall County, Texas. Center of the N½ of section 4, block H., J. H. Stephens Survey. Elevation, 2225 feet.

No. 6.—H C. Robinson *et al.* Kuteman No. 1, Donley County, Texas. 660 feet from the North and 330 feet from the West lines of SW¼ of section 88, block 20, H. & G. N. Survey. Elevation, 2704 feet.

DISCUSSION OF SECTION

Section A-B (Pl. 2) parallels the Red River syncline from Wichita County to Hall County, thence north to Donley County (Pl. 1). Boring No. 6 penetrated a rock column more or less characteristic of the succession indicated by borings flanking the eastern portion of the Amarillo-Wichita uplift in the Panhandle of Texas.

The base of the Wichita group in boring No. 1 is not so definite as it is farther south and west. Oil was found at a depth of 685 feet, and this may mark the base of the Wichita group. Where marine strata occur, as in boring No. 2, the base of the Leonard series is readily identified.

Section A-B shows that the base of the redbeds rises westward to Hardeman County, then drops in Hall and Donley counties. It is obvious that the base of the red color does not define a stratigraphic horizon, hence it can be used only locally for structural subsurface mapping. In the northwestern part of the Texas Panhandle, the Leonard has graded almost completely into redbed clastics. The presence of salt in borings Nos. 5 and 6 marks the easternmost recognized occurrence of salt. This salt horizon may be traced into the area from the north and west as it occurs in the Panhandle of Texas, western Oklahoma, and western Kansas. It occurs below the San Angelo horizon and its equivalents and may be correlated with the Stone Corral of Kansas and the Bullwagon of Texas. The Wellington salt of Kansas occupies the lower portion of the Leonard and probably in part is equivalent to the zone of *S. crassitectoria*. All salt beds in the Leonard are essentially free of potash, which distinguishes them from most of the salt deposits in the Custer group and its equivalents.

SECTION C-D

LOCATION OF BORINGS

No. 7.—S. I. Davis' C. C. Lynch No. 1, Throckmorton County, Texas. Center NE¼ of the NE¼ of section 157, B. B. B. & C. Survey. Elevation, 1605 feet.

No. 8.—Amerada Petroleum Corporation's C. J. Kleiner No. 1, Haskell County, Texas. 660 feet from the North and East lines of section 57, J. McCass Survey, A-95. Elevation, 1465 feet.

No. 9.—H. F. Wilcox Oil & Gas Company's W. F. Ross No. 1, King County, Texas. 330 feet from the North and West lines of section 23, R. B. Masterson Survey. Elevation, 1812 feet.

No. 10.—Humble Oil & Refining Company's Matador Land and Cattle Company No. 1, Dickens County, Texas. 1980 feet from the North and East lines of section 26, block 1, Shock and Arnold Survey. Elevation, 2477 feet.

No. 11.—Humble Oil & Refining Company's Matador Land and Cattle Company No. B-1. Motley County, Texas. 330 feet from the East and 7500 feet from the North lines of Blanco County School Lands, A-220. Elevation, 2662 feet.

DISCUSSION OF SECTION

Section C-D (Pl. 2), drawn at right angles to the regional strike (Pl. 1), shows several stratigraphic features of the Leonard. Boring No. 7 is a core test starting just below the Grape Creek beds of the Wichita group. It clearly shows the relation of the anhydrite zone to the base of the Leonard and underlying Elm Creek. This zone does not crop out due to solution by meteoric waters. Rainfall increases toward the east; hence anhydrites are better preserved in the surface expression of the Pease River group (Pl. 1).

The complete absence of redbeds in boring No. 11, except at the San Angelo horizon, is noteworthy. The strata from the base of the San Angelo to the base of the Leonard are almost all dolomite. The cherty dolomites above the San Angelo characterize this portion of the column over much of the Permian Basin but are not known to crop out on the east side. Boring No. 11 encountered thick beds of salt between 1330 and 2030 feet. They belong in the upper part of the Leonard far above the San Angelo horizon and are widespread in the northern portion of the Permian Basin, extending into Oklahoma and Kansas. The overlying Custer group also contains salt beds, but neither the lower nor upper salt beds crop out along the east side of the Permian Basin. The solution of several hundred feet of salt in this region simulates unconformities that are observable in the section above the San Angelo and especially in the several formations constituting the Custer group.

SECTION A-E

LOCATION OF BORINGS

No. 12.—British American Oil Company's J. E. Cope No. 1, Baylor County, Texas. 660 feet from the South and West lines of the NW¼ of section 220, T. & N. O. Survey, A-1369. Elevation, 1318 feet.

No. 13.—H. F. Wilcox Oil & Gas Company's J. E. Fritz No. 1, Baylor County, Texas. 3241 feet from the West and 1573 feet from the South lines of the NE corner of section 193, David Barlow Survey, A-10. Elevation, 1347 feet.

No. 14.—Merry Bros. & Perini, Inc. *et al.* C. L. Williams No. 1, Stonewall County, Texas. Center of SW¼ of SW¼ of section 6, block C., A. B. & M. Survey. Elevation, 1621 feet.

No. 15.—Humble Oil & Refining Company's Dallas Joint Stock Land Bank No. 1, Jones County, Texas. 4458 feet from the South and 990 feet from the West lines of section 267, M. Fragosa Survey, A-176. Elevation, 1804 feet.

No. 16.—Ajax Drilling Company's E. W Withers No. 1, Nolan County, Texas. 1320 feet from the South and West lines of section 65, block 21, T. P. R. R. Survey Elevation, 2323 feet.

No. 17.—L. G. Bradstreet's R. V. Kuteman No 1, Nolan County, Texas. Center of the NW¼ of section 95, block X., T. P. R. R. Survey. Elevation, 2488 feet.

No. 18.—Humble Oil & Refining Company's Lewis & Wardlaw No. 1, Tom Green County, Texas. 2080 feet from the South and 660 feet from the East lines of section 5, block 17, H. & T. C. Survey. Elevation, 2142 feet.

No. 19.—Stanolind's Williams No. 1, Irion County, Texas. 1320 feet from the East and South lines of section 1196, T. & T. R. R. Survey. Elevation, 2544 feet.

No. 20.—Big Lake's University No. 17-C, Reagan County, Texas. 2528 feet from the South and 2556 feet from the West lines of section 25, block 9, University Lands Survey. Elevation, 2678 feet.

DISCUSSION OF SECTION

Section A-E (Pl. 2) parallels the strike of Permian outcrops from Wichita County to Tom Green County (Pl. 1). Included is a composite type section of the outcrop on which are indicated the more commonly recognized geologic names. The section continues westward from Tom Green County to Reagan County.

Although fossils are lacking to the north and east yet correlations of the stratigraphic units in boring No. 18, Tom Green County, are relatively certain. So many wells have been drilled to the north and east that a well-to-well lithologic comparison brings out all essential details. Boring No. 18 starts on Comanchean beds. The top of the zone of *Perrinites* is at 250 feet. Regional mapping indicates that no Custer sediments intervene between the Comanchean and the zone of *Perrinites*. The base of the San Angelo was encountered at 605 feet. First fusulinid fragments were noted at 1200 feet. The zone of *S. crassitectoria* starts at 1800 feet, and the base of the zone of *Perrinites* is at 2410 feet. First fusulinids of the zone of *Properrinites* were noted at 2720 feet and are very abundant between 3000 and 3100 feet.

Boring No. 19 is not easy to correlate by lithology; accordingly paleontologic evidence must be sought. No single boring generally yields as much information concerning fossils as is necessary; hence studies on adjacent borings were used in making these correlations. Boring No. 19 starts on the Comanchean and at 650 feet enters the zone of *Waagenoceras*. No sediments of Custer age intervene between the base of the Comanchean and the zone of *Waagenoceras*. At 1230 feet *Parafusulina rothi* and *Parafusulina sellardsi* appear. These fusulinids are diagnostic of the lower and middle Delaware (Fig. 1). *Parafusulina* sp., which seems to be more highly developed than fusulinids in the underlying zone of *Perrinites*, has been found at 3030 feet. The top of the zone of *Perrinites* is placed at 3320 feet. The San Angelo probably is equivalent to the sandy zone between 3800 to 3850 feet. Fusulinids characterizing the zone of *Perrinites* occur between 4170 to 4200 feet. The top of the zone of *S. crassitectoria* is at 5300 feet, and the first fossils indicating the zone of *Properrinites* occur at 5920 feet.

Boring No. 20, in Reagan County, has the following section: Comanchean from the surface to 510 feet; Dockum to 730 feet; Elk City sandstone to 880 feet; base of Doxey shale 930 feet; base of the Whitehorse of Texas at 2930 feet. The base of the recently proposed Ochoa series is at a depth of about 1500 feet. In this boring sediments of Custer age rest upon the middle portion of the zone of *Waagenoceras*. Cores of strata between 3100 and 3200 feet are composed of *Parafusulina lineata*, *P. maleyi* var. *referta* and *P. rothi*, indicating a fusulinid bioherm. These fossils suggest that the strata between 2930 and 3900 feet are equivalent in age to the zone between 650 and 1800 feet in

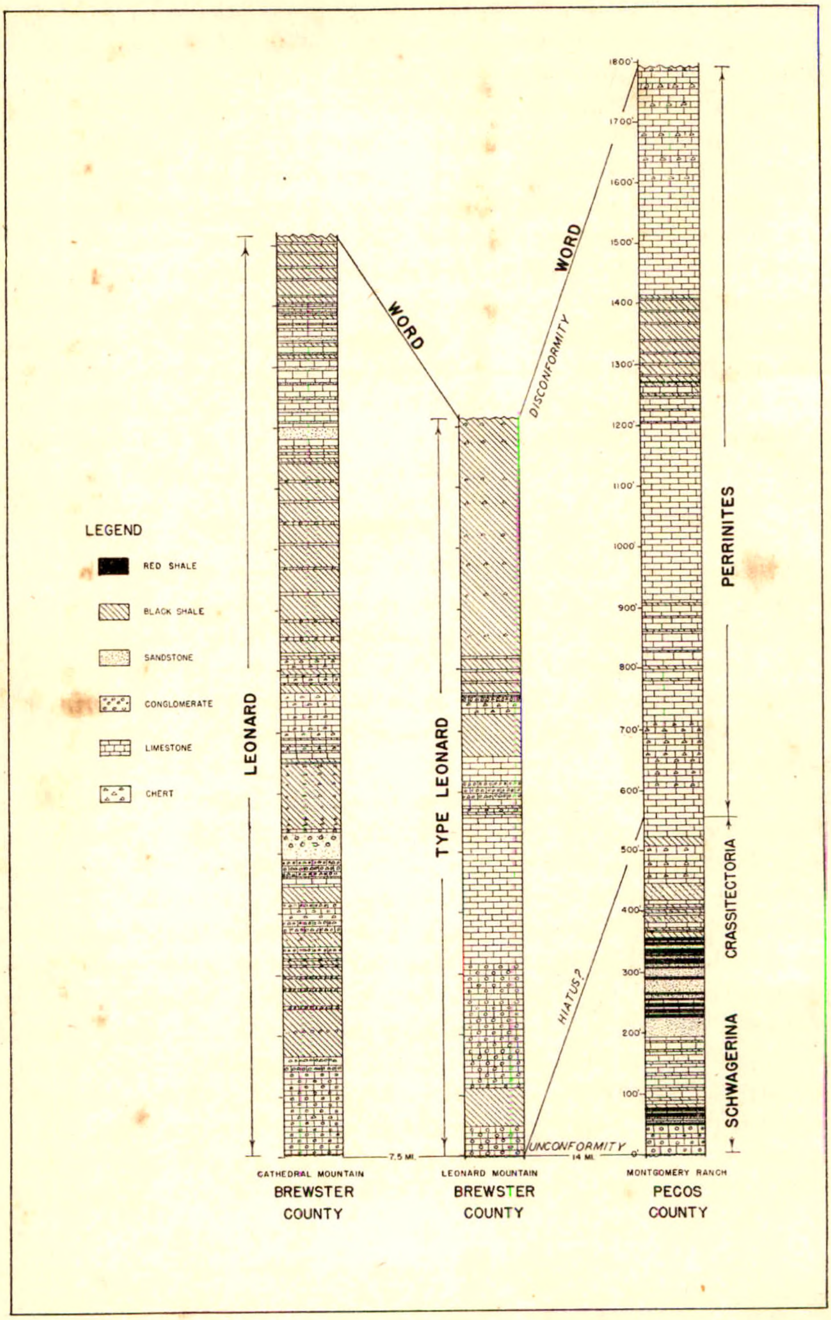

Figure 3. —*Cross section, type Leonard of Glass Mountains, West Texas*

boring No. 19. In boring No. 20, the zone of gritty gray-black shales gives way to black shales and cherty limestones at 4810 feet, which is considered the top of the zone of *Perrinites.* This contact is much the same as on the outcrop in the Glass Mountains. Determination of the base of the zone of *Perrinites* is problematical, but it is placed at 6290 feet, at which depth commercial oil production has been found in several wells in the Big Lake field. Inasmuch as oil has been found at or near unconformities, it is probable that this producing horizon marks the contact between the zones of *Perrinites* and *Properrinites.* Fusulinids found just below 6550 feet are too fragmentary to identify. If the above contact is correctly placed, the thickness of the zone of *Properrinites* is about 1700 feet. A number of borings, for which paleontologic evidence is available, have penetrated an equal or greater thickness of Wolfcamp beds. Since boring No. 20 is in the heart of the "black shale" basin, no subdivisions of the zone of *Perrinites* are attempted.

TYPE LEONARD OF GLASS MOUNTAINS, WEST TEXAS

Three measured sections (Fig. 3) of the outcrop at the type area of the Leonard in the Glass Mountains show the lithology of the *Perrinites* zone. Redbeds can be traced for several miles along the northeast part of the Glass Mountains scarp. They are wholly within the zone of *S. crassitectoria.* The indicated thickness of the zone of *Perrinites* in the type area is about half as much as it is along the east side of the Permian Basin.

The comparatively uninhabited nature of the Marathon region makes the location of geographic landmarks difficult. The general locations of the three measured sections used in the cross section (Fig. 3) are as follows: Montgomery Ranch section, Pecos County, is measured north-northwest from Montgomery Ranch in Hess Canyon Quadrangle, Marathon region. This section is 14 miles northeast of the Leonard Mountain section. The Type Leonard section, Brewster County, is measured north-northwest from the top of Leonard Mountain to Road Canyon, thence north to the summit of the north-dipping cuesta. It is 7½ miles northeast of Cathedral Mountain section, Hess Canyon quadrangle, Marathon region. Cathedral Mountain section, Brewster County, starts on the bolson flats of the Marathon plain and is measured N. 15° W. to the top of Sullivan Peak in the Cathedral Mountains, Altuda Quadrangle, Marathon region.

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RECONNAISSANCE SURVEY OF THE ROBERTS MOUNTAINS, NEVADA

BY C. W. MERRIAM AND C. A. ANDERSON

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ABSTRACT

The Roberts Mountains region, central Nevada, provides an excellent section of Paleozoic rocks ranging from Upper Cambrian to Permian Major low-angle thrusting is indicated by deformed Ordovician strata resting on Paleozoics of varying age Overlying a thick breccia zone, the upper thrust plate consists of sandstones, andesitic flows and tuffs, black shales, and bedded cherts (Vinini formation). Ordovician age of the Vinini is established on the basis of graptolite faunules A belt of Lower to Middle Ordovician graptolitic facies similar to the Vinini formation crosses the Great Basin west of Roberts Mountains Deposits of roughly the same age in the Roberts Mountains meridian and eastward are dominantly limestone, carrying distinct faunas. Axial planes of overturned folds in the thrust plate dip west, a further indication that the upper thrust plate moved from west to east. Minimum horizontal displacement is 16 miles. The date of thrusting is uncertain, but presumably was later Cretaceous or early Tertiary Following thrusting, an alaskite stock and rhyolite porphyry plugs were intruded; lava flows and tuffs covered the area in part. Thrust plate and cover of volcanics have been broken into normal fault blocks The post-thrusting igneous rocks, like volcanic rocks of Utah and New Mexico, are characterized by high potash content.

INTRODUCTION

Among the important results of recent geologic investigation in Nevada, western Utah, and southeastern California is recognition of widespread low-angle thrust faulting. Extending our knowledge of this type of deformation to a previously little-known area, a thrust plate of great magnitude is described in the Roberts Mountains of east-central Nevada. In addition the general geology of the area is presented briefly.

For several years Merriam has studied the Paleozoic stratigraphy and paleontology of central Nevada, emphasizing the Devonian system. Sections of the Devonian and adjacent rocks were measured by instrumental survey, and reconnaissance mapping was done to show distribution of the Devonian rocks. Some of the results of detailed stratigraphic work in the eastern portion of the Roberts Mountains quadrangle have been published (Merriam, 1940).

It soon became apparent that thrusting followed by normal faulting greatly complicated the structural picture, and that igneous activity, separating the two periods of deformation, required special attention. In order to clarify the geologic history of the region Anderson joined Merriam during the summer of 1939 in a continuation of the project. Reconnaissance mapping initiated by Merriam was carried forward, and Anderson spent some time on a study of the igneous rocks. In this paper, Merriam assumes responsibility for all statements relating to Paleozoic stratigraphy and paleontology; Anderson is responsible for the igneous rocks, and both writers have joined in discussion of the structure.

The Roberts Mountains, east-central Nevada, are located in the northeast corner of the quadrangle bearing this name, and about 25 miles northwest of Eureka; less than a quarter of the quadrangle has been mapped in this study. Fifteen miles south of the Roberts Mountains is the Lincoln Highway, serving as the southern boundary of the mapped area.

The quadrangle map (scale 1/250,000) was enlarged to twice original scale for field work, and the major formations and faults have been plotted. The writers wish to express their appreciation to the topographers for an excellent base map; the enlargement was found very serviceable in reconnaissance mapping.

ACKNOWLEDGMENTS

The writers are indebted to many geologists for advice and helpful criticism. Among these, special thanks are due Dr. Sheridan A. Berthiaume

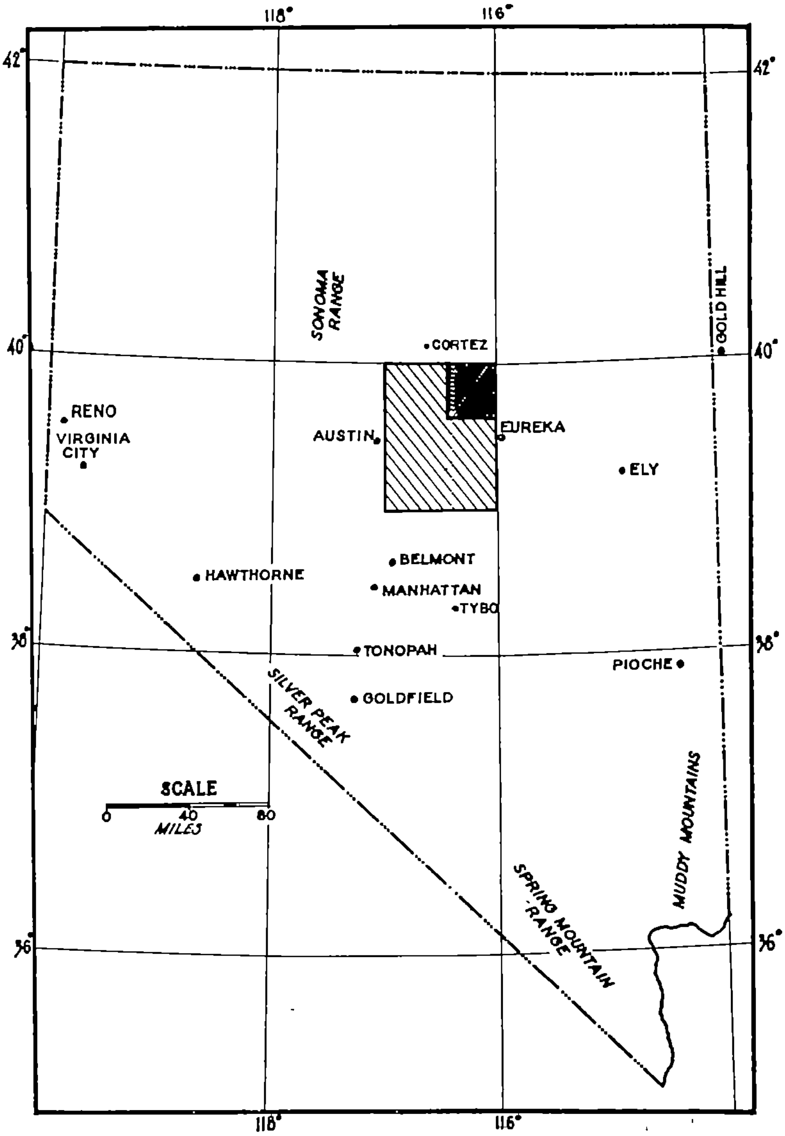

FIGURE 1.—*Index map of Nevada*

Roberts Mountains quadrangle shown in light shading; area of Plate 4 indicated by heavy shading. Localities referred to are also marked.

for help in the field. Berthiaume was investigating the Carboniferous in central Nevada and assisted in study of the later Paleozoic rocks of the Roberts Mountains region. Doctors T. B. Nolan and H. G. Ferguson of

the United States Geological Survey and Dr. S. W. Muller of Stanford University have been generous with advice, particularly valuable owing to their familiarity with the region. Special thanks are due Dr. Rudolf Ruedemann, whose graptolite identifications form a part of this study. Dr. C. M. Nevin of Cornell University and Dr. James Gilluly of the University of California obligingly read and criticized the manuscript. Grateful acknowledgment is made to The Geological Society of America for a grant from the Penrose Bequest used in support of field investigation and for chemical analyses of igneous rocks.

TOPOGRAPHY

The Roberts Mountains are roughly triangular in outline culminating in a central peak (Roberts Creek Mountain) with an elevation of 10,125 feet (Pl. 1, fig. 2). Western Peak, 4 miles northwest of Roberts Creek Mountain, rises abruptly from Denay Valley to an elevation of 9102 feet. Between Western Peak and Roberts Creek Mountain, a linear east-west ridge has two peaks each 9500 feet in elevation (Pl. 1, fig. 2); the prominent eastern eminence has been given the name Cooper Peak. Intermittent streams drain the mountain with a roughly radial pattern. One permanent stream, large for this region (Roberts Creek), flows southward from Roberts Creek Mountain (Pl. 1, fig. 1); another shorter permanent stream, Pete Hanson Creek, drains a small portion of the northwest part of the Roberts Mountains. Permanent springs are fairly numerous in the higher part of the range. The northern boundary is the broad alluviated Denay Valley, rising gently toward the southwest on the flank of a low divide separating Roberts Mountains from the Simpson Park Range. The western and southern boundaries are also alluviated forming the northern margin of Kobeh Valley. This is broken by a low ridge 6 miles long projecting south-southeast. To the east, the Roberts Mountains are separated in part from the Sulphur Spring Mountains by Garden Valley which dies out southward; the southeastern portion of the Roberts Mountains merges directly into the Sulphur Spring Mountains, interrupted by Mount Hope, an oval-shaped hill with summit elevation of 8412 feet (Pl. 2, fig. 3). The Sulphur Spring Mountains extend southward as a low north-south ridge culminating in Whistler Mountain with an elevation of 8167 feet. The alluviated margins of the Roberts and the Sulphur Spring Mountains range in elevation from 6000 to 6500 feet so that a maximum relief of 4000 feet is represented by Roberts Creek Mountain. Lone Mountain to the south of Roberts Mountains rises like an island above Kobeh Valley (Pl. 1, fig. 3); the summit with an elevation of 7940 feet is about 1800 feet above its base.

GENERAL STRATIGRAPHY

Paleozoic strata occupy the surface over some three-fourths of the Roberts Mountains area. Lavas and locally stratified tuffaceous deposits cover much of the remaining part. These extrusive igneous rocks are presumably Cenozoic. In addition several classes of post-Paleozoic intrusives have been recognized. The oldest exposed Paleozoic rocks are Upper Cambrian, while the latest beds of this era have been assigned to the Permian (Table 1). Mesozoic rocks are not known, though in the adjoining Eureka district terrestrial Cretaceous beds have been reported (Nolan, oral communication; MacNeil, 1939). Possible Triassic strata are present in the Sulphur Spring Mountains, though of such occurrence there is no confirmation (S. A. Berthiaume, oral communication). Owing to faulting and lack of outcrop a portion of the later Devonian and earlier Carboniferous does not appear at the surface in the Roberts Mountains proper but is represented in the southern portion of the mapped area at Lone Mountain and Devils Gate. Here and there on the lower alluviated slopes bordering Kobeh Valley are exposures of white sands and silts representing Pleistocene lake beds; in some localities as at Lone Mountain these contain abundant diatoms.

Two genetically distinct successions of Paleozoic sediments are recognized in the Roberts Mountains region: one regarded as the normal section, the other representing sediments of a different belt. The last owe their present location to thrust fault displacement. Strata of the thrust plate are of Lower and Middle Ordovician age, while those of the normal section extend from Upper Cambrian to Permian.

NORMAL SEQUENCE OF PALEOZOIC STRATA

GENERAL STATEMENT

The normal sequence is well exposed in the central Roberts Mountains (Pl. 1, fig. 2), at Lone Mountain (Pl. 1, fig. 3), in the Sulphur Spring Mountains, and at the southern end of the Whistler Range. In the central Roberts Mountains the section is fairly continuous from the base of the Eureka quartzite (Middle Ordovician) to the upper Nevada formation (Middle Devonian), while at Lone Mountain there appears to be no important physical break between the upper Pogonip (Chazyan) and the middle of the Devils Gate formation (higher Middle Devonian). In both sections the sedimentation is, excepting the Eureka quartzite, predominantly calcareous with minor amounts of argillaceous debris. In the Roberts Mountains the maximum unbroken section of Paleozoics is of the order of 8500 feet; at Lone Mountain the entire section exposed measures 7300 feet. The later Paleozoics are well exposed in the Sulphur

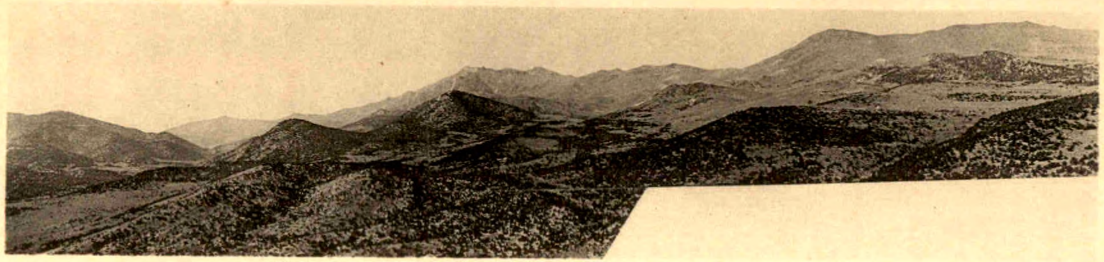

FIGURE 1. LOOKING NORTH TOWARD ROBERTS CREEK MOUNTAIN
Lava-capped mesa to right. Prominent hills in middleground center and left, Nevada limestone.

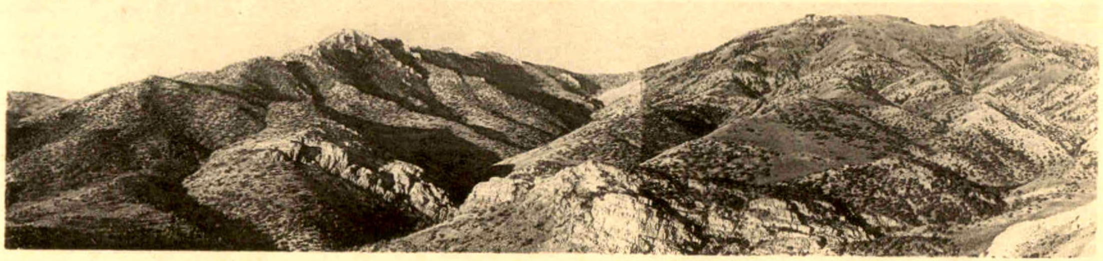

FIGURE 2. LOOKING EAST TOWARD COOPER PEAK (LEFT) AND ROBERTS CREEK MOUNTAIN (RIGHT)
Easterly dipping Paleozoic section in fault contact to the north (left) with dolomites of Cooper Peak. Prominent cliff in foreground Eureka quartzite.

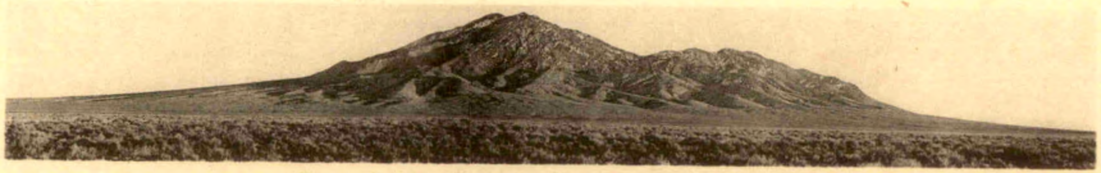

FIGURE 3. LONE MOUNTAIN VIEWED FROM THE NORTHWEST
Homoclinal block of Paleozoic sediments dipping eastward (to the left), ranging from Pogonip Ordovician (right base) to Devils Gate Devonian at top of section.

ROBERTS MOUNTAINS

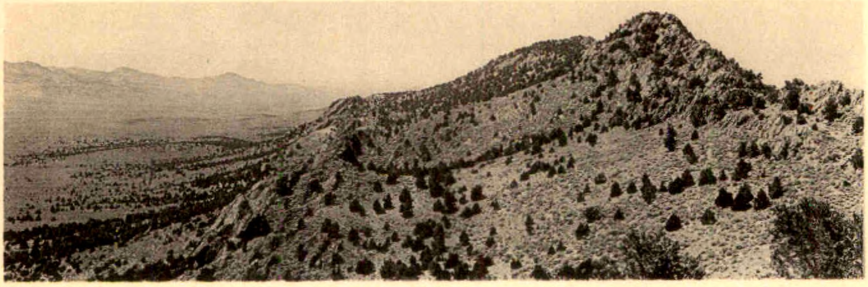

FIGURE 1. UNDIFFERENTIATED CARBONIFEROUS CHERT-PEBBLE CONGLOMERATES AND QUARTZITES DIPPING STEEPLY WESTWARD (OVERTURNED)
Looking north.

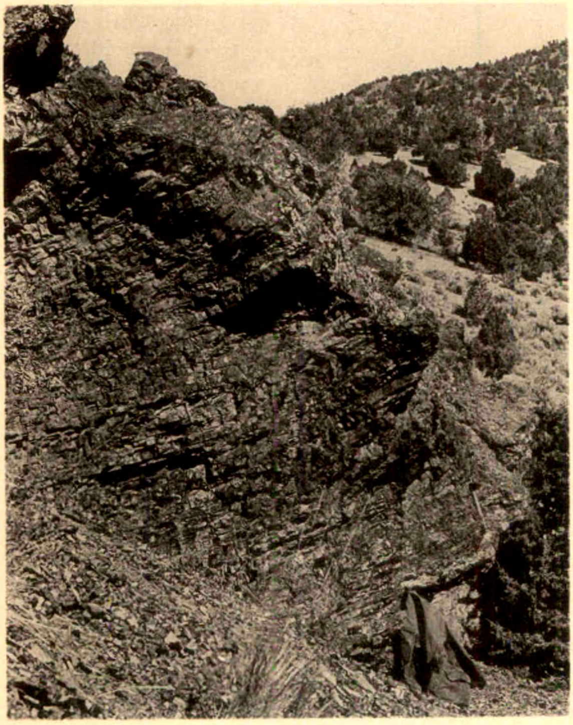

FIGURE 2. BEDDED UPPER VININI CHERTS WITH BLACK SHALE INTERBEDS
Exposed along Dry Creek.

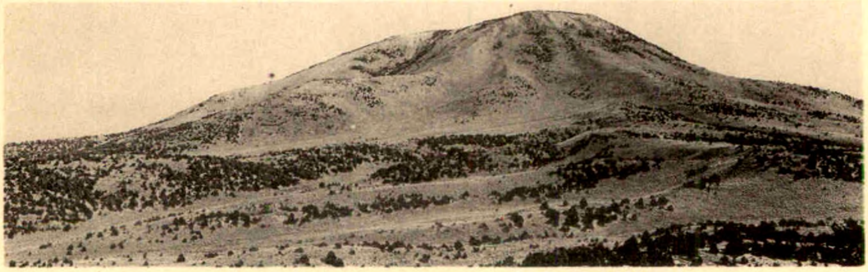

FIGURE 3. LOOKING EAST TOWARD MOUNT HOPE
Plug of rhyolite porphyry intruded in the Vinini sediments.

UNDIFFERENTIATED CARBONIFEROUS, VININI CHERTS AND MOUNT HOPE PLUG

TABLE 1.—*Normal Section in Roberts Mountains Region*

Period		Formation	Description
PERMIAN CARBONIFEROUS		Undifferentiated Permian and Pennsylvanian in Sulphur Spring Mountains 5000 feet	Reddish conglomerate with limestone cobbles; chert pebble conglomerate and quartzite; lower part with sandy to pebbly limestones and fusulinid limestones
		(discontinuity, greater part of Diamond Peak unrecognized)	
		Diamond Peak beds	Black clay shales, brownish sandstones and conglomerates
DEVONIAN	Upper	Devils Gate formation 2065 feet	Dark-gray well-bedded limestone and calcareous shale; lower part heavily bedded or massive light- and dark-gray to black dolomitic limestones
	Middle Lower	Nevada formation 2400 feet	Massive or heavily bedded gray and black limestones, siliceous limestones and dolomitic limestones in upper part; lower part dark-gray well-bedded shaly or argillaceous limestones
SILURIAN		Lone Mountain formation 2200 feet	Massive and blocky dolomites of light-gray smoke-gray and dark-gray color
		Roberts Mountains formation 1900 feet	Well-bedded dark slate-gray limestones, crystalline limestones; dark cherty limestone member at base
ORDOVICIAN	Upper	Hanson Creek formation 560 feet	Dark-gray heavily bedded limestone, thin-bedded bluish-gray limestone; dark-gray dolomitic limestone below
	Middle	Eureka quartzite 500 feet	Upper 300 feet mostly white vitreous quartzite; lower 200 feet with dark- and light-gray quartzite predominant
		(discontinuity in Roberts Mountains, Pogonip group, upper part unrecognized)	
CAMBRIAN	Lower	Pogonip group, lower part; undifferentiated Lower Ordovician and Upper Cambrian	Dark gray thin-bedded limestones; dolomites

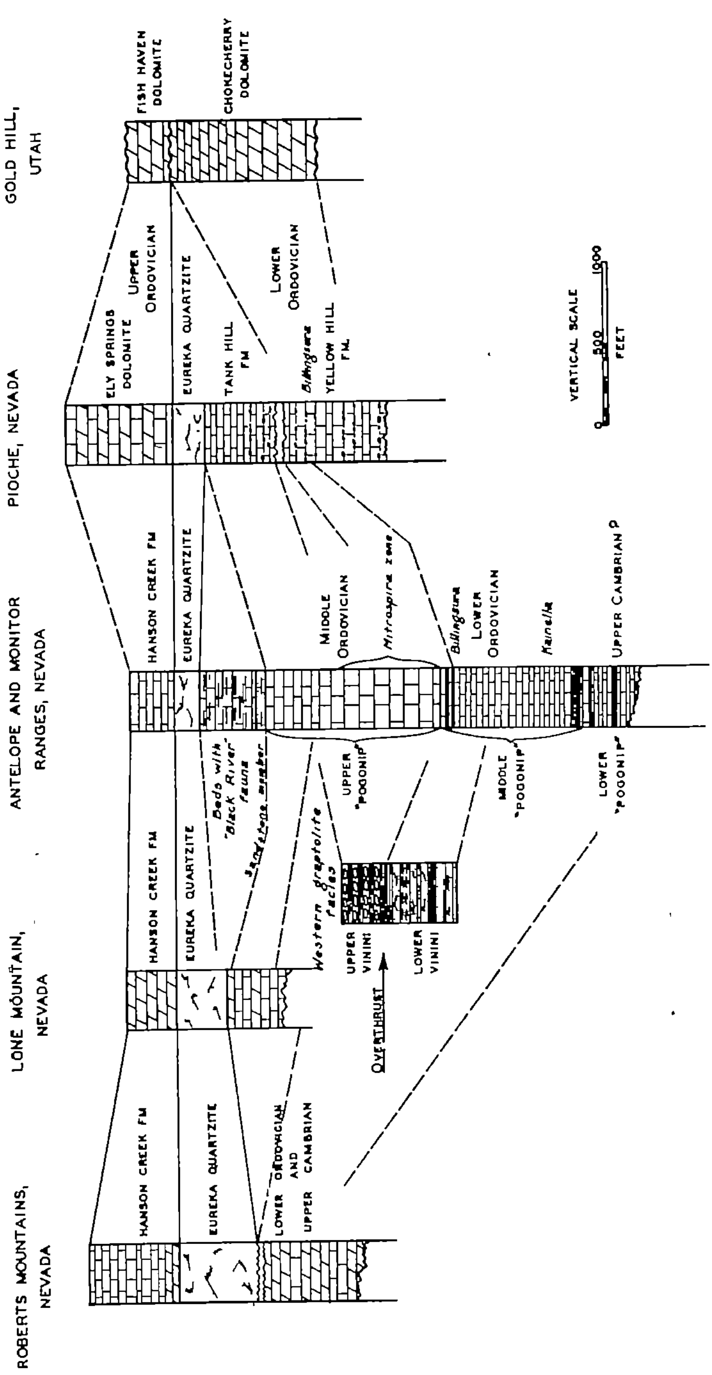

FIGURE 2.—*Correlation chart of the Ordovician*

Spring Mountains to the east, where Carboniferous and Permian lie in fault contact with older rocks. To the south in the Whistler Range and at Devils Gate are the most complete sections of higher Devonian and the only definitely known exposures of Lower Carboniferous in the area covered by the present survey.

POGONIP GROUP

The name "Pogonip limestone" has been used broadly for strata in the Great Basin lying between the base of the Eureka quartzite and the top of the Upper Cambrian Dunderberg shale (Eureka district). Following this interpretation, the "Pogonip" may embrace faunal zones ranging from Upper Cambrian to approximately Black River Middle Ordovician. Recent studies in the Antelope and Monitor ranges 35 miles south of Roberts Mountains indicate within the vertical limits of the "Pogonip limestone" in this region at least three recognizable formational units, each with distinctive faunal characterization. Walcott (1923, p. 466) proposed the name "Goodwin formation"[1] for the lower 1500 feet of the Pogonip in the Eureka district. If accepted, the term "Goodwin formation" requires more specific lithologic and faunal delimitation, for the lower 1500 feet of the Pogonip as defined by Hague includes important stratigraphic boundaries and at least two distinct faunas. In the Antelope-Monitor area (Fig. 2) the lowermost Pogonip appears to be Cambrian, while the succeeding Pogonip with *Kainella* is correlative with "Ozarkian".

Within the Roberts Mountains the Pogonip group is poorly represented. Where best developed on the west flank of Western Peak the rocks below the Eureka quartzite show little similarity to the thick Pogonip section of the Antelope and Monitor ranges 50 miles to the south. The Roberts Mountains Pogonip includes thinly bedded dark-gray limestone and dolomites showing heavier bedding. Local evidences of replacement indicate that the dolomite is in part secondary. No evidence was found of the upper Pogonip limestones which in the Monitor Range are over 1000 feet thick. Fossils collected at Western Peak within 500 feet of the Eureka quartzite appear to be Upper Cambrian. The fossils are as follows: *Eurekia* cf. *granulosa* Walcott, *Agnostus* sp., and *Acrotreta* sp. In the section on the west side of Roberts Creek Mountain Kirk (1933, p. 31) found fossils some 300 feet below the Eureka quartzite, which are "of lower Pogonip age (approximately Beekmantown) and rather early in that". As recognized by Kirk there is an apparent gap of some significance with much, perhaps the greater part, of the Pogonip group absent. Whether erosion or nondeposition is the correct explanation has not been determined. Cutting out of the higher divisions of the Pogonip group either by normal faulting or thrusting does not seem possible.

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[1] The "Goodwin" bears a fauna which Walcott assigned to the "Lower Ozarkian" Ulrich and Cooper (1938, p. 24, 26) regard the "Goodwin" as Upper Ozarkian.

The section at Cortez, 18 miles north-northwest of Roberts Creek Mountain, shows about 350 feet of Eureka quartzite underlain by roughly 1000 feet of barren dark-gray dolomites and magnesian limestones which bear little resemblance to Pogonip of the better-known sections to the south. In absence of fossils it is a reasonable assumption that all the pre-Eureka sediments represented at Cortez may be Cambrian.

The character and age of the deposits immediately below the Eureka quartzite change southward from Cortez and Roberts Mountains; at Lone Mountain Chazyan upper Pogonip limestones are found below the Eureka quartzite, while in the Antelope-Monitor area beds of approximately Black River age occur beneath the main body of quartzite, and above Chazyan upper Pogonip.

EUREKA QUARTZITE

Exposures of Eureka quartzite are found on the northwest side of Roberts Creek Mountain, on the western flank of Western Peak, and at Lone Mountain. Judging from the difference between thicknesses in the Roberts Mountains and at Lone Mountain the Eureka apparently thins toward the south; the greater decrease in the southern Antelope valley area, 20 miles south of Lone Mountain, further substantiates this.

The Eureka on the northwest side of Roberts Creek Mountain is about 500 feet thick. As noted by Kirk (1933, p. 30) the basal third is sharply differentiated from the higher portion, being much darker and occasionally showing cross-bedding. The upper 350 feet has the dense vitreous or gleaming-white character typical of the formation in many areas. Evidences of bedding are usually not recognizable in this portion of the formation. Toward the top, the formation tends to lose much of its dense, vitreous character and is terminated upward by a patchy calcareous sandstone phase (Merriam, 1940, p. 11).

At Lone Mountain the Eureka is 350 feet thick, about 150 feet less than in the Roberts Mountains. It is similar to the same unit at Roberts Creek Mountain with the lower portion in general darker than the upper, and occasionally showing cross-bedding. According to Kirk (1933, p. 30-31) some 50 feet of brownish-weathering sandy dolomite lying between the "typical uppermost Pogonip limestone" (Chazyan) and the cliff-forming Eureka apparently pertains to the latter. Kirk has suggested the possibility that these lower sandy deposits together with the darker, occasionally cross-bedded lower unit of the Eureka may be equivalent to the combined *Endoceras*-bearing sand member and overlying Black River deposits of the Antelope and Monitor ranges (Fig. 2). According to this interpretation southward thinning of the quartzite deposits may be accounted for in part by lateral replacement of the normally arenaceous lower division of the typical Eureka by a partly calcareous facies as seen in the Antelope and Monitor ranges. No apparent lack of conformity between the Eureka and underlying strata was observed at Lone Mountain and southward in the Antelope Valley region.

Hague (1892, p. 56-58) recognized an unconformity between the Eureka and overlying "Lone Mountain limestone" (unrestricted) of the Eureka district. Evidences of denudation prior to deposition of the "Lone Mountain limestone" are cited by Hague with reference to sections in the mountains connecting the Fish Creek Range with Prospect Ridge, while in other parts of the Eureka district different horizons of the "Lone Mountain limestone" and even of the Devonian are said to lie directly upon the quartzite. However, the clearest evidence of unconformity at this position is given by Ferguson (1933, p. 20-21) for the Tybo area, 80 miles south of Eureka. Here Silurian dolomites and limestones rest upon an eroded and channeled surface of Eureka quartzite; where the quartzite is missing they lie directly on Pogonip limestone. The lower or Ordovician part of the "Lone Mountain limestone" in the older sense is not represented, possibly removed by erosion.

Within the area of the present survey no convincing physical evidence of post-Eureka stripping or channeling has been recognized, and faunal evidence is still too scanty to establish erosion or nondeposition of portions of the higher Ordovician section. In view of the nature of the lithologic change passing from the faunally barren quartzitic deposits of the Eureka to the calcareous and distinctly marine beds of the later Ordovician it seems reasonable to expect here a lost interval of some magnitude insofar as the purely marine record of the Ordovician is concerned. As no fossils have been found within the Eureka quartzite proper the possibility that it may be in part at least of continental origin has not been eliminated.

HANSON CREEK FORMATION

The Upper Ordovician Hanson Creek formation rests with probable disconformity upon the Eureka quartzite. The type section of the formation is on the northwest side of Roberts Creek Mountain and crosses the east fork of Pete Hanson Creek. Hague's (1892, p. 57) "Lone Mountain limestone" has been subdivided by Merriam (1940, p. 13) with the Hanson Creek formation representing the lower or "Trenton" portion. Thin calcareous sandstones, locally patchy, lying above the typical vitreous Eureka quartzite have been regarded as the basal member of the Hanson Creek. Above appear dark-gray dolomitic limestones found in place 10 to 15 feet above unquestionable vitreous Eureka quartzite. The lower dolomitic division is not over 40 feet thick at the type section. The remainder of the formation is medium-gray, gray, and bluish-gray limestone most of which is fine-grained. Much of the lower half of the formation is thin-bedded and slabby or shaly limestone bearing a *Cryptolithus* fauna apparently corresponding with that listed by Hague (1892, p. 59) from the "Trenton" portion of the lower "Lone Mountain limestone" northeast of Wood Cone, Eureka district. The total thickness of Hanson Creek formation at the type section is 560 feet. It is overlain conformably by the basal chert member of the Roberts Mountains Silurian.

At Lone Mountain the Hanson Creek formation is only 318 feet thick and is largely medium- to dark-gray dolomite and dolomitic limestone. Streptelasmoid corals occur at the base, but otherwise identifiable fossils are rare. In this respect the section differs decidedly from that on Pete Hanson Creek. In the Monitor Range, 20 miles south of Lone Mountain, the Hanson Creek formation is largely well bedded flaggy limestone and calcareous shale. Here several zones throughout the formation contain graptolites. The following forms have been identified by Ruedemann from the Hanson Creek beds of the Monitor Range:

Climacograptus tridentatus Lapworth var. *maximus* Decker
Orthograptus calcaratus Lapworth var *trifidus* Gurley
Orthograptus sp.
Dicellograptus complanatus Lapworth var. *ornatus* Lapworth

According to Ruedemann *Dicellograptus complanatus* var. *ornatus* occurs on Trail Creek in the Hailey quadrangle, Idaho, while the other species occur in the Lower Sylvan beds of Oklahoma and the Polk Creek beds in Arkansas where *Dicellograptus complanatus* also is found. Decker (1935, p. 698) considers the Sylvan to represent upper Richmond at the top of the Ordovician. Otherwise the rather large fauna of the Hanson Creek remains unstudied. In the Monitor Range the Hanson Creek beds are overlain by the Roberts Mountains formation bearing abundant *Monograptus*.

ROBERTS MOUNTAINS FORMATION

The type section of the Roberts Mountains formation lies on the west side of Roberts Creek Mountain where it is 1900 feet thick. The lower member is a siliceous limestone, with lenses and interbeds of bluish-black chert. This cherty zone is very persistent in central Nevada. The upper limit of the formation is the base of light-gray dolomites of the Lone Mountain formation (restricted). The Roberts Mountains formation or its equivalents formed part of the original "Lone Mountain limestone" of

Hague. At the type section the formation consists largely of dark slate-gray limestones with only a moderate amount of dolomite. This contrasts sharply with the overlying Lone Mountain formation (restricted) which is almost entirely dolomitic. Much of the formation exhibits slabby or flaggy parting. The lower and medial portions include many crinoidal beds. In the upper 800 feet the bedding tends to become more massive, while the upper 220 feet of the formation is partly dolomitic. The relationship with the overlying Lone Mountain formation is apparently gradational. Fossils are numerous throughout the Roberts Mountains formation. Most characteristic are pentameroid brachiopods of the genus *Conchidium* or *Virgiana.* Corals are abundant with *Halysites,* Heliolitidae, favositids, and small solitary rugose forms within the lower 300 feet. Many horizons throughout the remainder of the formation have yielded excellent coral assemblages, including Heliolitidae, many forms of tabulates, and several compound Rugosa which suggest *Strombodes.*

At Lone Mountain the Roberts Mountains formation, including the basal chert member, is only 741 feet thick and largely dolomitic. Fossils are relatively rare. In the Monitor Range the Roberts Mountains interval is represented by the basal cherty limestone succession and overlying well-bedded limestones.

Two species of *Monograptus* have been identified by Ruedemann from a horizon a short distance above the cherty member in the Monitor Range; *Monograptus acus* Lapworth and *Monograptus pandus* Lapworth. According to Ruedemann (written communication)

> "*Monograptus acus* was already known to the writer from the Trail Creek beds in the Hailey quadrangle, Idaho; *M. pandus* had already been listed by me from the Brisco limestone in British Columbia and was further observed in collections from Galena Creek, tributary of Prairie River, N. W. T Also two varieties are noted here from Trail Creek."

As pointed out by Ruedemann *M. acus* is listed by Elles and Wood from British zones 23 and 24 and *M. pandus* from zones 22, 23, and 24. On this basis the lowermost zone of the Roberts Mountains formation which yielded the graptolites is "of the approximate age of the lower and middle Gala beds of Great Britain, or of Clinton (and younger age) of New York."

In the higher beds of the Monitor Range a large and well-preserved Silurian fauna has been found; this assemblage appears to represent a facies quite different from any recognized in the Roberts Mountains. A few of the corals of the type section of the Roberts Mountains formation have been recognized in the Monitor Silurian strata and also at Ikes Canyon in the Toquima Range. Evidently Silurian strata older than the Lone Mountain formation (restricted) are widespread in central Nevada.

LONE MOUNTAIN FORMATION (RESTRICTED)

The name Lone Mountain formation has been redefined and restricted by Merriam (1940, p. 13). The "Lone Mountain limestone" of Hague (1892, p. 57) included that part of the Paleozoic section at Lone Mountain and in the Eureka district lying between the Eureka quartzite and the Nevada limestone of Devonian age. In the original usage the formation included at least three definable lithologic units and embraced Upper Ordovician, Silurian, and possibly Lower Devonian strata. At the type section the Lone Mountain as restricted is 1570 feet thick, while in the Roberts Mountains it increases to 2200 feet. The formation consists of light mouse-gray and darker smoky-gray dolomites and dolomitic limestones usually weathering to light mouse gray. Bedding is generally obscure or massive, and the outcrop surfaces are frequently blocky. At many points the contact relationship of the Lone Mountain with the underlying Roberts Mountains Silurian and with the overlying Nevada limestone of Devonian age appears to be gradational. At Lone Mountain the basal contact is within a zone where the dolomites change from the predominantly dark gray of the Roberts Mountains formation to lighter grays of the higher Silurian dolomites. On the west side of Roberts Creek Mountain a bed of smoke-gray crinoidal dolomite with crinoid stems half an inch in diameter lies near the base of the Lone Mountain formation.

Below the main body of the Nevada formation in the Roberts Mountains and in the Sulphur Spring Mountains a zone of dolomitic limestone bearing a distinctive Lower Devonian fauna is found. *Spirifer* cf. *arenosa* is common. Though sometimes similar lithologically to the Lone Mountain dolomite these deposits are on a faunal basis referred to the Nevada as a basal member

Other than a questionable *Syringopora* found near the base of the unit at Lone Mountain no determinable fossils have been obtained from the Lone Mountain formation. However, the age is reasonably well fixed by overlying and underlying formations as Upper Silurian with the possibility that some portion of the early Devonian may also be represented. A study of the large faunas from the Roberts Mountains formation at the type locality and Monitor Range should place more exact limitations on the age of the Lone Mountain, in view of Lower Devonian assignment of the overlying beds with *Spirifer* cf. *arenosa.*

NEVADA FORMATION (RESTRICTED)

The term Nevada formation has been redefined (Merriam, 1940. p. 14) to apply only to the lower part of the "Nevada limestone" of Hague (1892, p. 63). Merriam designated the upper portion of the original "Nevada limestone" as Devils Gate formation. The type section of the Nevada formation as redefined lies at Modoc Peak in the western part of the original Eureka sheet (Hague, 1892). The formation is calcareous throughout and largely limestone, but in the upper 600 feet, as at Lone Mountain, many dolomitic members occur.

The subzone bearing *Spirifer* cf. *arenosa* is included with the Nevada formation as its basal division; lithologically it is sometimes similar to the underlying Lone Mountain. Above the lower division 500 to 700 feet of the Nevada may be distinguished by lithology from the higher members. In general this portion is inclined to be dark gray, thinly bedded, with argillaceous or shaly partings, and rich in fossils. The upper two-thirds of the formation is more heavily bedded and includes siliceous limestones and dolomites. In general the upper members are sparingly fossiliferous as compared with the lower, though certain zones in the silicified portion (*Martinia kirki* zone) yield silicified fossils in abundance, while toward the top of the formation deposits of stromatoporoid limestone and *Cladopora* limestone are common. The uppermost members of the Nevada fall in the *Stringocephalus* zone. Deposits of this zone are frequently medium-gray to dark-gray and black dolomites or magnesian limestones. These beds often contain abundant fragmentary silicified shells of *Stringocephalus*. The upper limit of the Nevada formation (restricted) is established at the top of the highest bed containing *Stringocephalus* in view of the abundance, broad geographic distribution, and limited vertical range of this important brachiopod genus. The Nevada formation has been subdivided (Merriam, 1940, p. 9) into four major paleontologic zones with a number of subzones and horizons recognized on fossil evidence. The major zones are as follows from bottom to top: (1) *Spirifer kobehana* zone; *Spirifer pinyonensis* zone; (3) *Martinia kirki* zone; (4) *Stringocephalus* zone.

In the Roberts Mountains the highest Nevada beds recognized belong in the *Martinia kirki* zone, for the uppermost Nevada or *Stringocephalus* zone and the Devils Gate formation apparently were cut out by faulting or occupy brecciated belts within which stratigraphic determination is not possible. The section at Lone Mountain is probably the best in central Nevada for study of the Nevada formation. In the Sulphur Spring Mountains on the eastern border, the lower Nevada is well exposed. Here are the best-known occurrences of the Lower Devonian *Spirifer* cf. *arenosa* beds which lie immediately east of a major normal fault separating Permian or Upper Carboniferous sediments from those of Lower Devonian and upper Lone Mountain (Silurian) age.

DEVILS GATE FORMATION

In redefining and subdividing the "Nevada limestone" of Hague the Devonian beds of central Nevada which lie above the *Stringocephalus* zone were designated as the Devils Gate formation (Merriam, 1940, p. 16). The type section is at Devils Gate where about 1100 feet of dark-gray limestones and subordinate calcareous shales is exposed. Only about half the formation is exposed at Devils Gate. However, to the south and east along Modoc Ridge lower and lower strata appear until at Modoc Peak (original Eureka sheet) an 1800-foot sequence of the Devils Gate formation reveals the basal members. At this point the upper 265 feet is not exposed. The total thickness of the Devils Gate in the type area is 2065 feet. The upper limit is marked by disappearance of the *Cyrtospirifer* or "*Spirifer disjunctus*" fauna, and the dark, bluish-gray, highly calcareous shales pass into platy, black, and usually noncalcareous shales of the Diamond Peak Carboniferous. The lower 900 feet consists largely of thick-bedded, light-gray, and very dark-gray dolomites and dolomitic limestones including numerous members filled with stromatoporoids and *Cladopora* Other organic remains are rare in this portion of the section. The upper part of the formation as exposed at Devils Gate is largely limestone and much more fossiliferous, particularly the uppermost 800 feet. Within the upper 100 feet occurs a bed of edgewise limestone conglomerate showing angular and flat fragments of dark laminated limestone in a matrix of lighter argillaceous lime. The abundant fossils in this bed seem to occur almost entirely in the darker reworked fragments and not in the matrix.

The Devils Gate has been subdivided faunally (Merriam, 1940, p. 9) into three zones, from top to bottom as follows: (1) *Cyrtospirifer* zone; (2) *Phillipsastraea* zone; (3) *Spirifer argentarius* zone. The lowest fossiliferous beds of the *Spirifer argentarius* zone recognized lie from 800 to 900 feet above the base, defined by the highest known occurrence of *Stringocephalus*.

Distribution of the Devils Gate formation is very limited within the area of the present survey. Other than the type section the only known outcrop is at Lone Mountain where the lower 1100 feet is exposed. In the Roberts Mountains these beds appear to have been cut out by faulting or so badly brecciated as to be unrecognizable.

DIAMOND PEAK BEDS

Lower Carboniferous strata representing the Diamond Peak are found only in a small area at Devils Gate, resting upon the *Cyrtospirifer* beds of the Devils Gate Devonian. The deposits are dark, usually noncalcareous shales with interbedded thin, impure quartzitic layers and occasional lenses of fine pebble conglomerate. A sill of alaskite porphyry has invaded the contact zone of the Devils Gate and Diamond Peak where these strata are best exposed on the north side of Devils Gate pass. In this vicinity the overthrust Vinini black shales are also present. When graptolites are absent it has been found difficult to separate these Ordovician deposits from the Diamond Peak beds which resemble them lithologically.

The best sections of the Diamond Peak beds are found at Diamond Peak and Newark Mountain on the east side of the Eureka district. Here the black Lower Carboniferous shales at the base are separated from the Devils Gate Devonian by a black crinoidal limestone, probably pertaining to the former.

While no true Diamond Peak beds have been recognized with certainty in the Roberts Mountains it is possible that a few small black shale areas here mapped as Vinini Ordovician and others included with undifferentiated Paleozoic, may represent windows of this Carboniferous division.

The term Diamond Peak beds is employed to include not only the original "Diamond Peak quartzite" of Diamond Peak (Eureka district) but in addition the underlying black shales and sandy interbeds referred by Hague (1892, p. 156-157) to the "White Pine shale". Actually the "Diamond Peak quartzite" includes a very large percentage of conglomerate at Diamond Peak, and the lithologic term is misleading. As the stratigraphic relations of the true White Pine formation in the White Pine district are not well understood it appears undesirable to use this term for dark shales of the Eureka and Roberts Mountains areas.

UNDIFFERENTIATED CARBONIFEROUS AND PERMIAN

A belt of Carboniferous and Permian rocks outcrops conspicuously in the western portion of the Sulphur Spring Mountains. Standing nearly vertical the resistant conglomerate and quartzites of this succession produce a line of strike ridges which may be followed for more than 16 miles. The ridges are particularly prominent north and south of the mouth of Tyrone Creek. In their southern extension these rocks generally dip eastward at very high angles; in the middle part the beds are locally overturned, dipping steeply to the west (Pl. 2, fig. 1). The Carboniferous and Permian rocks are in fault contact on both east and west sides of the belt. On the east a major fault brings the lower Nevada and the Lone Mountain dolomite into contact with Permian for a distance of 11 miles. While to westward the Ordovician Vinini is normally faulted against Carboniferous, it appears likely that the Vinini was previously thrust over these younger rocks. The later normal faults are assumed to cut the thrust plate giving the present visible relationship (Pl. 4).

The lowest Carboniferous beds exposed are limestones which have yielded a fusulinid faunule. Certain of the limestones in this portion of the section are argillaceous, showing mud cracks.' Above this facies sand grains and pebbles are admixed with the limestone, and the bedding is frequently characterized by cross-lamination. The ridge-forming conglomerates follow. Within this portion of the section at least two recognizable subdivisions were found: (1) a lower very resistant chert pebble conglomerate including some interbeds of sandstone or quartzite; (2) a higher conglomerate containing many limestone cobbles and pebbles. In the upper conglomerate the cementing material is commonly red; red sandy and silty lenses are present, particularly near the top of the succession. Locally the lower or chert pebble conglomerate shows evidence of post-depositional silicification and sometimes appears almost homogeneous, with occasional faint outlines of the original chert pebbles. At present no attempt is made to subdivide these rocks on a paleontologic basis.[2] However, the lowest fusulinid limestones are in all probability Pennsylvanian, while the later reddish conglomerates and silts are regarded as Permian.

DOLOMITES OF COOPER PEAK

The east-west trending ridge culminating in Cooper Peak consists of fractured, veined, and altered limestones and dolomites (Merriam, 1940, p. 35). Little evidence of bedding is visible, and the eastern portion of the dolomite is similar to that of the Lone Mountain formation. Two and a half miles west of Cooper Peak a large and well-preserved fauna of silicified brachiopods and corals was found in badly fractured unstratified dolomitic limestone. The assemblage, including *Trematospira cooperi* and *Spirifer* cf. *arenosa,* is Lower Devonian. In the Sulphur Spring Mountains this fauna occurs below the shaly division of the Nevada limestone and below the range of several fossil forms characteristic of the upper *Spirifer kobehana* zone. The dolomite is separated from underlying formations by a low-angle thrust fault dipping to the west. The fault is well defined 350 feet below the summit of Cooper Peak. Directly below the thrust fault to the east and north of Cooper Peak lower and middle Nevada limestones dip about 45° to the southwest with middle Nevada below the lower Nevada indicating overturning of beds beneath the thrust. At other points north and northwest of Cooper Peak evidence of overturning of lower Nevada beds below the thrust plane was found. Probably the dolomites of the Cooper Peak mass include much disturbed rocks ranging from pre-Lone Mountain Silurian to lower Nevada Devonian.

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[2] Dr S A Berthiaume has completed a detailed study of the Carboniferous and Permian rocks of the Roberts Mountains region, the Eureka district, and the Diamond Range Presentation of the results of stratigraphic investigation in the Sulphur Spring Mountains is therefore not attempted in this reconnaissance study

UNDIFFERENTIATED PALEOZOIC ROCKS

An area of dark shales along Jackass Creek north of the conspicuous Nevada limestone window shows lithologic features suggesting that it may not pertain to the Vinini overthrust plate. In spite of lithologic similarity to the Vinini no graptolites were found, and the sediments contain abundant peculiar subcylindrical bodies several inches long which have somewhat the appearance of badly preserved plant (lycopod) stems. Whether these pieces are actually organic or wholly concretionary has not been determined. Possibly this is a window exposing sediments as late as Carboniferous. While the occurrence of Diamond Peak Lower Carboniferous black shale beneath the overriding thrust plate has not clearly been demonstrated, it is probable that it exists here.

ROCKS OF ROBERTS MOUNTAINS THRUST PLATE

GENERAL STATEMENT

Lower and Middle Ordovician strata differing lithologically and faunally from the calcareous Ordovician rocks of the normal section outcrop over a wider area in the Roberts Mountains than those of any other Paleozoic division. They represent a plate thrust at a low angle over the underlying rocks. Since there are excellent exposures and good fossil localities along Vinini Creek the name Vinini formation has been applied to these strata.

Rocks of the Vinini formation have been traced from McClusky Pass on the west to the Sulphur Spring Mountains on the east, a distance of 20 miles. In a north-south direction the outcrop of Vinini extends practically unbroken from the mouth of Dry Creek to the region east of Yahoo Canyon in the Eureka district, a distance of 30 miles. Outliers and erosion remnants indicate a former more extended distribution.

The Vinini formation actually embraces at least two and possibly three formational divisions. Owing to complexities of structure and small scale of the base map it has not been feasible to cartographically illustrate a subdivision. When large-scale maps are available the Vinini can probably be separated into units which will bring out many features of the structure not dealt with in this report. For the present two divisions, lower Vinini and upper Vinini, are recognized.

The exposed normal section does not appear to include stratigraphic equivalents of the Vinini formation in the Roberts Mountains, owing to absence of the upper, and probably much of the middle, Pogonip as a result of pre-Eureka erosion or nondeposition. In the Antelope and Monitor ranges, however, the Pogonip includes probable stratigraphic but not facies or faunal equivalents of the entire Vinini formation (Fig. 2).

LOWER VININI

General statement.—Lower strata of the Vinini formation are best exposed on the east side of Roberts Creek Mountain, along the main east fork of Vinini Creek. The deposits here consist of dark-gray, brownish-weathering bedded quartzites, gray arenaceous limestones or calcareous sandstones commonly showing cross-lamination, and fine laminated sandy and brownish-gray and greenish-brown silty sediments. Locally the silty beds have a shaly parting, and in some localities true black shales are present in the lower Vinini. Within the lower Vinini no attempt has been made to work out in detail the vertical order of various subsidiary lithologic units owing to the folding and overturning of these beds.

Lava flows and tuffs.—Near the top of the lower Vinini, lava flows and tuffs occur at several localities. These are particularly well shown about 3 miles south of Roberts Creek Mountain just east of Roberts Creek. Here 40 feet of tuff-breccia grades upward into several feet of greenish volcanic sand overlain by cherty shale, typical of the upper Vinini. Nearby an amygdaloidal lava flow about 40 feet thick is exposed, also overlain by cherty shale of the upper Vinini. At two other localities, only amygdaloidal lava was observed, such as the outcrop previously noted along Vinini Creek and on the west flank of Roberts Creek Mountain 4 miles east of the mouth of Jackass Creek. The lava flow along Roberts Creek has numerous thin tabular calcic andesine phenocrysts imbedded in a finely crystalline groundmass of albite microlites, slender iron oxide needles presumably pseudomorphic after amphibole, rare chlorite crystals, and epidote granules. Secondary calcite is present both as amygdaloidal fillings and veinlets. In the other exposures of lava, the plagioclase phenocrysts consist of turbid albite, in part replaced by calcite, associated with a groundmass of albite microlites, chlorite flakes, and iron-ore granules. In some amygdules, fibrous chlorite surrounds a calcite core. One specimen of lava contains no feldspar phenocrysts, but pseudomorphs of chlorite and granular quartz preserve the outlines of former hornblende phenocrysts. The matrix of the tuff-breccia contains large and small lithic fragments similar to the lavas described above. Crystal fragments are very turbid albite and aggregates of chlorite frequently spherulitic. Calcite and granular quartz are the cementing material. The tuffs and lavas presumably were hornblende andesites that have been chloritized, and, with the exception of the lava flow containing andesine phenocrysts, extensive albitization has taken place.

Turner (1902) reported, in the Silver Peak Range, 170 miles to the south-southwest, rhyolitic and dacitic tuffs and lavas interbedded with dark cherts of the Palmetto group, which is also Ordovician. However, the Palmetto igneous rocks are said to be interbedded with "Normanskill" cherts and therefore on the whole may postdate the Vinini extrusives occurring at the top of the lower Vinini of Deepkill or Beekmantown age, and below the upper Vinini cherts. Igneous rocks have not definitely been found within the main body of upper Vinini cherts which probably correlate roughly with the Normanskill Palmetto cherts.

Age of lower Vinini.—Among the abundant graptolites of the lower Vinini on Vinini Creek, Ruedemann has recognized the following forms:

Dictyonema n. sp.
Tetragraptus similis (Hall)
T. quadribrachiatus (Hall)
Phyllograptus cf. *angustifolius* (Hall)
Didymograptus nitidus (Hall)
Isograptus gibberulus Nicholson (*Didymograptus caduceus* Salter)
Cardiograptus folium Ruedemann

In addition small orthoid brachiopods and the trilobite *Pliomerops* occur in sandy or quartzitic layers immediately above the graptolite faunule listed. Hexactinellid sponge spicules are very abundant in certain graptolite beds.

Ruedemann regards the graptolite assemblage as clearly of the same age as the *Tetragraptus* beds of the lower Deepkill shale in New York State, or of lower Canadian (Beekmantown) age. The fauna is also well represented in the Ordovician of Idaho and British Columbia. Of particular interest is the Australian genus *Cardiograptus* found also in the Idaho and British Columbia faunules.

Several additional graptolite localities have been found in the lower Vinini on both the east and the west side of the Roberts Mountains as well as the central part of the range. In some places the graptolites occur in quartzitic beds as well as in the silty and shaly deposits. Judging from diversity of the graptolite faunules at these localities probably several graptolite horizons are represented within the lower Vinini.

UPPER VININI

General statement.—A succession of bedded cherts and black organic shales constitutes the upper Vinini which follows the lavas and tuffs of the lower division. In most sections studied the chert beds and lenses make up more of the thickness than the shale interbeds, though at certain exposures only the shale was recognized. Except where the resistant cherts are exposed, outcrops of this unit are often poor (Pl. 3, fig. 3).

The chert layers generally vary in thickness from 1 to 4 inches (Pl. 2, fig. 2), while the organic shale partings are frequently less than 1 inch thick. In spite of the black color of the organic shale and the black to dark-gray and greenish color of much of the chert when fresh, both shale and chert tend to weather limonite brown, cream, white, and grayish white. From a distance, whitish outcrops of upper Vinini may be mistaken for tuff. Along Vinini Creek and at other localities throughout the Roberts Mountains region the shales are readily ignited. The oil yield on distillation of selected samples is above 25 gallons per ton. Where the black shale interbeds have a high content of bituminous matter, the dark chert layers probably carry much of the same.

Within the shale interbeds rounded or walnut-shaped nodules are common. These sometimes consist of a combination of siliceous and bituminous material with some calcium carbonate and often much pyrite. The nodules resemble small coal balls and on sectioning reveal well-preserved organic remains (Merriam and Daugherty, 1938). The shales and particularly the nodules are highly graptolitic, also bearing large quantities of Radiolaria, and microscopic algae. The highly bituminous or oil-bearing shale facies is apparently limited to the central and eastern Roberts Mountains belt, extending from the Vinini Creek drainage southward to Devils Gate.

Upper Vinini cherts.—The siliceous strata are clearly not the result of silicification of earlier sediments but are original primary bedded cherts, separated by thin shale partings. The chalcedonic cherts are pale green to black, with layers ranging from 1 to 4 inches in thickness, a few reaching 1 foot or more. On a fresh moistened fracture surface, clear round areas can be detected with a hand lens, resembling outlines of Radiolaria such as characterize cherts of the Jurassic Franciscan formation in the Coast Ranges of California. Under the microscope, these areas appear as clear minute aggregates of quartz or chalcedony, the latter arranged in spherulitic fibers, similar to the Radiolaria as described by Davis (1918, p. 257) in the Franciscan cherts In some, the chalcedonic fibers are surrounded by microgranular quartz. The clear areas are surrounded by cryptocrystalline silica which polarizes feebly. In the gray to black cherts, dark-brown isotropic spots are visible in thin section. These resemble organic matter (bituminous?) such as that which characterizes some Miocene cherts of the Coast Range in California (Davis, 1918, p. 288). Quartz veinlets are common in the cherts representing minute fracture lines later sealed with quartz.

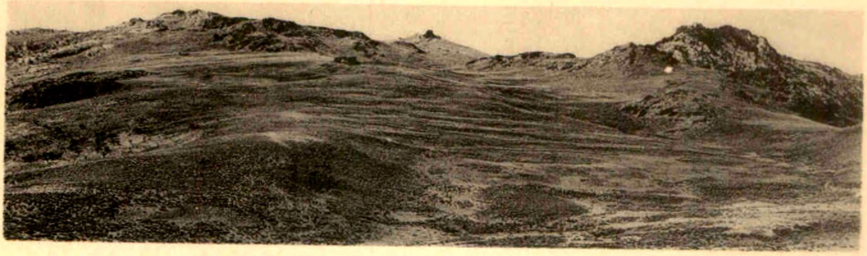

FIGURE 1. EAST SIDE OF ROBERTS CREEK MOUNTAIN
Illustrating crags (middleground) of eroded, brecciated base of upper thrust plate. White crag to right is composed of Nevada (?) limestone. Foreground shows rocks of Vinini formation.

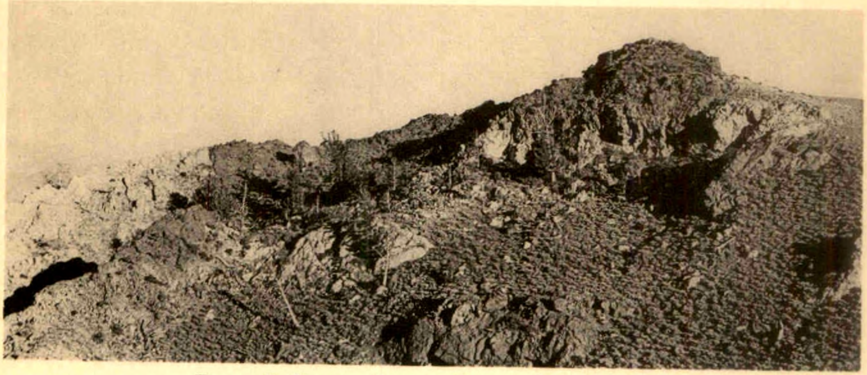

FIGURE 2. BRECCIA ZONE AT BASE OF THRUST PLATE
East slope of Roberts Creek Mountain.

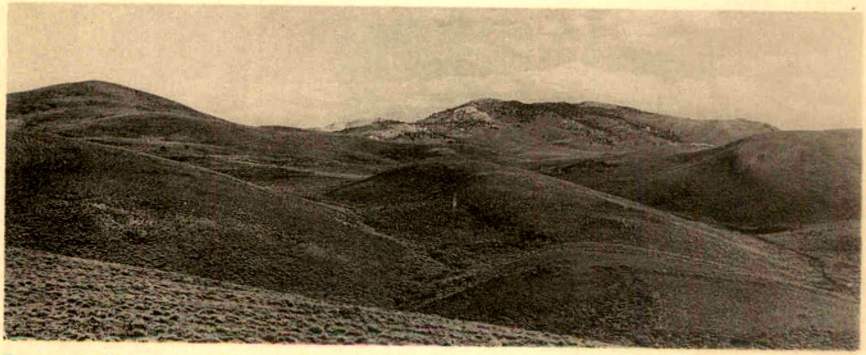

FIGURE 3. WINDOW OF NEVADA LIMESTONE (BACKGROUND) PROJECTING THROUGH OVERTHRUST VININI
Southeast of Jackass Creek. Rounded topography in foreground characteristic of much of the Vinini.

ROBERTS MOUNTAINS THRUST PLATE

The occurrence of chert above lava as noted on Vinini Creek is interesting in view of the suggestion that the siliceous sediments owe their origin directly or indirectly to volcanic activity (Taliaferro, 1933). In this connection attention is called to the work of Ruedemann and Wilson (1936) who have suggested a relationship between the widespread occurrence of Normanskill radiolarian cherts in eastern North America and an episode of igneous activity. The petrographic character and manner of occurrence of the New York Normanskill cherts is quite similar to the upper Vinini with, in both cases, association of cherts and the pure type of graptolite shale (Ruedemann, 1935). However, in the eastern New York area chert is most abundant in the lower part of the Normanskill, while the upper Vinini appears to represent upper Normanskill.

Age of upper Vinini.—From black organic shales of this division the following graptolite species have been identified by Ruedemann:

Leptograptus flaccidus (Hall) var *spinifer* Elles and Wood mut. *trentonensis* Ruedemann
Dicranograptus spinifer Lapworth
Diplograptus angustifolius (Hall)
Orthograptus calcaratus var. *acutus* Lapworth
Climacograptus bicornis (Hall), narrow variety
C. modestus Ruedemann
Retiograptus geinitzianus (Hall)

In addition to graptolites, the upper Vinini beds contain abundant Radiolaria, thin-shelled protremate brachiopods, and microscopic protophycean algae (Merriam and Daugherty, 1938). Ruedemann has concluded that this faunule is of upper Normanskill age and identical with faunules from Bruno Creek, Bay Horse quadrangle, Idaho, Fall Creek, Hailey quadrangle, Idaho, and British Columbia. Typical *Dicranograptus spinifer* as found in the Vinini shale has not been observed in eastern North America but occurs in the Idaho localities here mentioned.

Garden Pass (Summit) graptolite zone.—An important graptolite faunule was collected at Garden Pass (Summit), Nevada, by Walcott (Gurley, 1896, p. 302). This assemblage, one of the first found in western America, occurs on the old right of way of the defunct Eureka and Nevada railroad where it traverses the low pass separating the Roberts Mountains block from the Sulphur Spring Mountains. The exact association has not been recognized elsewhere in the Vinini formation and probably represents a zone between that of the lower Vinini and the upper Vinini organic shales and cherts. On a paleontologic basis it is therefore likely that at least a three-fold subdivision of the Vinini formation is possible with a time range from lower Canadian to upper Chazyan. The long interval of Ordovician time involved emphasizes the possible thickness of these highly dislocated and eroded overthrust rocks.

Graptolites collected by Walcott (Gurley, 1896, p. 302) "at the crossing of the Eureka and Palisade Railroad at Summit, Nevada," have been identified by Lapworth, Gurley, and Ruedemann. The forms listed by Gurley are as follows:

Dendrograptus cf. *serpens* Hopkinson
Didymograptus perflexus Gurley
Phyllograptus sp. indet.
Diplograptus sp.
Climacograptus caelatus
Cryptograptus tricornis
Glossograptus ciliatus
Thamnograptus anna

Ruedemann regards the form referred to *Glossograptus ciliatus* by Lapworth (manuscript report) and by Gurley (1896) as a "mutation" of this species which he described as *Glossograptus ciliatus* mutation *horridus*. According to Ruedemann (1908, p. 384) the following forms are associated at the Garden Pass (Summit) locality:

Glossograptus ciliatus mutation *horridus*
Phyllograptus anna
Climacograptus caelatus
Cryptograptus tricornis
Caryocaris

On the basis of this assemblage the deposits at Garden Pass (Summit) have been regarded by Ruedemann as probably Upper Beekmantown.

Whether the graptolite specimens collected by Walcott and identified by Lapworth and Gurley were all obtained from exactly the same zone is not known. The possibility that they represent more than one zone must be considered in view of structural complexity in these shaly deposits at Garden Pass (Summit). The graptolitic deposits in this region are known to include several clearly defined divisions. As listed the Garden Pass (Summit) faunule is distinct from those recognized on Vinini Creek and at other localities throughout the mountain area, appearing on paleontologic grounds to require a stratigraphic position between the lower Vinini and upper Vinini assemblages.

Because of profound dislocation of the Vinini formation it has not been possible to obtain reliable figures for thickness of the two divisions. However, on Vinini Creek the combined thickness remaining probably cannot be less than 500 feet, and much of this section has certainly been removed by erosion. Black graptolitic shales and cherts at the north end of the Simpson Park Range west of McClusky Pass appear to represent only the upper Vinini and may greatly exceed the figure given for this division alone.

FACIES CHANGE IN LOWER AND MIDDLE ORDOVICIAN

GENERAL STATEMENT

West of the Roberts Mountains throughout a broad belt extending 170 miles south-southwest from the Simpson Park Mountains and Toquima Range to the Silver Peak Mountains of western Nevada, marked and complex facies changes are to be expected in rocks of Lower and Middle Ordovician (Pogonip) age. Kirk (1933, p. 29) in his description of the Eureka quartzite notes that west of the 117th meridian "The Middle Ordovician is represented largely by graptolite shales with minor amounts of sandstones, limestones, and calcareous shaly beds". The 117th meridian is the western border of the Roberts Mountains quadrangle but lies 30 miles west of the Roberts Mountains proper. All available data indicate that east of this variable belt the deposits of Canadian (Beekmantown) and Chazyan age are predominantly marine limestones or dolomites with subordinate arenaceous and argillaceous facies. Westward the known Lower and Middle Ordovician includes a high percentage of argillaceous and arenaceous strata with a notable amount of graptolitic shale and chert. With respect to the Roberts Mountains and the Antelope-Monitor area the terms western and eastern facies may conveniently be employed in a general sense.

EASTERN FACIES

Deposits of the eastern facies are not well shown in the Roberts Mountains proper where, as discussed above, only the lower part of the Pogonip has been recognized. However, in the same meridional belt at Lone Mountain and in the Antelope-Monitor area 20 miles south of this locality, typical representation of the eastern facies is found. While the Lone Mountain exposures show only the upper or Chazyan portion to advantage, the Antelope Monitor sections below the Eureka quartzite include some 2300 feet of strata extending from Early Ordovician or "Ozarkian" to the Middle Ordovician Black River beds (Fig. 2). In these more southerly sections limestone predominates, with subordinate fine sandy and argillaceous deposits. No Vinini-type rocks or faunas are known to occur in the normal stratigraphic sequence.

WESTERN FACIES

A great development of black graptolitic shales and cherts is found in the Simpson Park Range adjoining the Roberts Mountains on the west, but it is not known whether these represent a westward extension of the Roberts Mountains upper thrust plate or are part of an unthrust sequence.

West of Potts in the poorly known northern part of the Toquima Range true graptolite shales of Ordovician age have not been identified, though graptolites occur sparingly here in the Pogonip limestones and abundantly in adjacent Silurian calcareous beds. In the southern part of this range, however, the Ordovician section includes much argillaceous and arenaceous sediment, suggesting partial replacement of the Pogonip limestones by these clastic facies.

Graptolite shales have long been known at Belmont, some 85 miles south-southwest of Roberts Mountains on the eastern margin of the Toquima Range (Gilbert, 1875, p. 180). At this locality the beds have been subjected to varying degrees of metamorphism. In the Manhattan district near the south end of the Toquima Range and at Belmont, Ferguson (1924, p. 20-25) has described the Ordovician Toquima formation. The division is commonly marked by a basal quartzite which nowhere exceeds 50 feet and in places is lacking. Above this are dark slates with Normanskill graptolites. Below the Toquima formation are schists, schistose slates, and limestones which on a basis of lithologic similarity to beds above the Normanskill graptolites are tentatively assigned to the Ordovician. Above the graptolitic slate of the Toquima is a gray limestone with jaspilite identical with the limestone below the Toquima. While above the Toquima limestones no constant succession of beds could be recognized, the bulk of superjacent Toquima apparently consists of chloritic schist with beds of dark siliceous slate or chert and a few intercalations of crystalline limestone. Interbedded and interfingering quartzite and graptolitic slate are also found within the complex but in uncertain position. Broadly speaking, lithologic similarity to the Vinini formation is noteworthy, though no detailed comparison is yet possible.

In the Silver Peak Range, 170 miles south-southwest of Roberts Mountains, Turner (1902) described the Palmetto formation as consisting of dark thin-bedded cherts with layers of gray graptolite shales, smaller amounts of reddish slates, and occasional limestone layers. Two graptolite horizons occur, the lower containing graptolites of Beekmantown age and the higher with a Normanskill faunule. Reported igneous rocks associated with the cherts have been discussed above. Again there exists a similarity to the Vinini formation.

Fifty miles west of Roberts Mountains in the Toyabe Range, Emmons (1870, p. 324) noted a quartzite series overlain by slates "which included fissile limestone shales, more or less siliceous clay-slates, and locally, schistose and somewhat crystalline rocks, resembling mica and hornblende schists, and in one instance, a marbleized limestone." Spurr (1903, p. 94-95) has suggested that these slates are probably the same as the slates of the Toquima Range to the south, and on the basis of graptolites found in that range by Gilbert (1875, p. 180) the rocks in the Toyabe Range were dated by Spurr as lower Silurian (Ordovician in the present usage). These strata have been intruded by granites and have undergone some metamorphism, thus differing from the Vinini formation. The same situation has been recognized in the Toquima Range (Ferguson, 1924, p. 38) where the Ordovician rocks are intensely folded and locally metamorphosed.

Correlation of the western graptolitic rocks with the normal Pogonip section in the Antelope-Monitor region is unsatisfactory. However, the upper Vinini with its Normanskill graptolite faunas probably represents some portion of the Chazyan upper Pogonip interval (Fig. 2), though a correlation with the superjacent "Black River" beds is conceivable. The lower Vinini with Deepkill graptolite faunas probably falls within the limits of the Canadian portion of the Pogonip near the zone carrying *Billingsura*. Occurrence of *Phyllograptus* cf. *loringi* in the *Billingsura* zone (Kirk, 1934, p. 455) of the Antelope Range is significant.

Though information concerning Lower and Middle Ordovician rocks in western Nevada is scanty,[3] all facts point to the existence of one or perhaps several more or less lineal marine belts or subordinate troughs in which graptolitic facies are prominent. The belts apparently extended north-northeast across the Great Basin from the Silver Peak area, communicating with central Idaho where correlative graptolitic beds are found. Faunal evidence further suggests a connection with British Columbia. Comparison with conditions in the restricted Ordovician faunal belts of the Appalachian geosyncline may be fruitful at a later date.

ROBERTS MOUNTAINS THRUST

GENERAL STATEMENT

In terms of structural geology the most striking feature of the Roberts Mountains is a very extensive low-angle thrust resulting in discordant position of Vinini Ordovician strata on Paleozoic rocks of varying age. Independent minor thrusts are also recognizable. Breaking of a local fold has been suggested to explain the origin of that beneath the Cooper Peak dolomite mass.

Some of the smaller folds in this area are intimately related to the major thrusting, while the broad major flexures are older, though possibly pertaining to an earlier phase of the same diastrophic episode. In general the folds of the Roberts Mountains region are of four types: (1) broad flexures affecting Paleozoic rocks below the thrust surface; (2) isoclinal folds within the major thrust plate; (3) minor irregular folds within the major thrust plate; (4) minor drag folds associated with normal faults.

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[3] Additional thick exposures of western facies Ordovician are known in the southwestern part of the Hawthorne quadrangle, Miller Mountain and Candelaria Hills, and northern part of the Sonoma Range quadrangle (H. G. Ferguson, personal communication)

On limbs of the broad flexures the strata vary in strike from north-south to northwest or northeast, while observed dips range from 25° to over 50°, with overturning in exceptional cases. At Lone Mountain and Roberts Creek Mountain there is a gradual but on the whole marked increase in dip within the same section passing from the lower (Ordovician) to the higher (Devonian) strata (Merriam, 1940, p 19, 33). The broad folds have been extensively eroded and usually appear only as homoclines (Lone Mountain) which represent but a portion of one limb. Closure is rarely encountered in these structures. Where crests or keels may exist much fracturing and minor faulting is to be expected, and the beds rarely carry over from one limb to another in continuity.

Isoclinal folding is recognized locally and is inferred to be very common within the thrust sheet. Widespread minor irregular folds in the thrust sheet resulted, like the isoclinal folds, from forces causing the thrust movement. Axes of the irregular folds vary greatly and were probably controlled to a large extent by irregularities in the surface over which the thrust sheet moved. Overturning is common

The more important normal faults and fault zones within the mountain mass are of two different sets with respect to direction: those which trend from north-northwest to slightly east of north, and those extending transversely in more or less an east-west direction. Displacement on faults of the first set may exceed 6000 feet as on Pete Hanson Creek where upper Nevada has been carried downward in contact with Eureka quartzite and older rocks. Most of the normal faults recognized are later than the major thrusting.

EVIDENCE OF MAJOR THRUSTING

Evidence for the thrust is greatly strengthened by occurrence of Ordovician graptolites at numerous scattered localities (Pl. 4). Resemblance of the lower Vinini with dark quartzitic members to the Carboniferous Diamond Peak beds was at first highly confusing. In the absence of faunal evidence at the outset of this study it seemed a natural conclusion that these deformed strata were younger than the Nevada Devonian, merely representing local thrust plates.

Mapping brings out lack of accordance of the Vinini formation to structure of the underlying rocks. Where the base of the thrust plate is exposed, it is usually marked by a very pronounced breccia zone, at least 30 to 40 feet thick (Pl. 3, figs. 1, 2). These breccias are best developed where the cherty layers of the upper Vinini have been fragmented. On the east slope of Roberts Creek Mountain near the summit area, chert-breccias grade upward into bedded cherts. Thin sections of the chert-breccias reveal radiolarian tests in the fragments with no recrystallization of the cryptocrystalline base. They are strongly cemented by quartz. The resistant thrust breccias are often silicified and stained by iron oxide so that they outcrop locally as brownish-black crags. No slickensided surfaces were found at the base of the thrust plate or in the underlying rocks. A revealing exposure near the summit area of Roberts Creek Mountain shows an intimate mixture of chert and limestone fragments at the base with "dikes" of breccia 4 feet wide and 12 feet long injected in massive limestone. In other exposures, near the summit, chert-breccia from the overlying Vinini rests upon limestone-breccia derived from the underlying Nevada formation.

The thrust surface is highly irregular with windows of underlying rocks now exposed by erosion (Pl. 3, fig. 3). Most windows which yielded fossils are of the Nevada formation. On the eastern flank of Roberts Creek Mountain the fault surface dips eastward from 5° to 20°. On the western flank of the mountain, the surface is inclined to the west at a moderate angle, but no exposures give an exact measure of the dip.

The Vinini beds are unmetamorphosed but have been intensely folded and crumpled with strikes ranging from N. 30° W. to N. 40° E.; in the main these trends appear to average about N. 10° W. The dips are also variable but in general range from 30° to 50°. In spite of moderate dips, overturning is often observed, the axial planes dipping west. An exception to this can be seen around Whistler Mountain where vertical dips and isoclinal folds are common. Possibly the muscovite alaskite stock comprising Whistler Mountain caused this local tight folding by pushing aside the Vinini formation at the time of intrusion.

DIRECTION OF MOVEMENT

In the following discussion, the term "direction of movement" of the upper thrust plate is used in a relative sense with no implication that only the upper block moved. It is believed that much more mapping must be done in the surrounding region before sufficient evidence is available to suggest whether "overthrusting" or "underthrusting" dominated.

Dip of the thrust surface does not indicate direction of movement, for both east and west dips are widespread (Pl. 4). Inclination of axial planes in small overturned folds within the Vinini is probably more significant, as in most instances the axial planes were found to dip westward.

Eastward movement of the Vinini plate is therefore indicated; this necessarily presupposes development of the small folds at the time of thrusting.

Perhaps the most convincing evidence for eastward movement is seen in distribution and facies relations of Great Basin Ordovician deposits Rocks and faunas of the Vinini formation are totally unlike those of normal Pogonip in the eastern facies, showing on the other hand decided similarity to Ordovician of the western belt. Strata of the upper plate are not, therefore, thought to be of local origin but to have moved from the west where the Ordovician is known to include variable sediments with predominant graptolitic facies.

No conclusive evidence regarding total movement of the upper plate is available. However, in the Roberts Mountains the Vinini thrust rocks are exposed almost continuously for an east-west width of 16 miles (Pl. 4), giving this minimum value of horizontal displacement. Undoubtedly the true horizontal movement is much greater. Whether the black graptolitic shales and cherts of the adjoining Simpson Park Range to the west have also undergone horizontal displacement is not known.

Determination of actual thrust magnitude must await more detailed lateral tracing of the two distinct Ordovician facies. So far, data on intergradation or rate of change from western to eastern facies are lacking. The possibility that the two were in no way directly connected at the time of deposition must also be considered As pointed out by Ferguson (personal communication), the eastern facies at Tybo, Hot Creek Range, is only about 25 miles in an air line from the western facies at Belmont, Toquima Range. In absence of evidence for horizontal movement between these two points, the distance seems short for such complete change in lithology.

Whereas it would be premature to comment extensively upon the relation of the Roberts Mountains thrust to other known major thrusts in Nevada and adjoining regions, it is significant that several of these give evidence of eastward movement. In the nearby Eureka district Nolan (personal communication) recognizes important thrusting with displacement in this direction. Among those more remote is the Muddy Mountains thrust (Longwell, 1922) some 260 miles south-southeast, where the upper plate moved from the southwest. Thrusts described by Glock (1929), Nolan (1929), and Hewett (1931) in the Spring Mountains west of the Muddy Mountains show relative eastward movement of the overlying block.

STRUCTURAL HISTORY

In unraveling the complex structural history, the following suggestions are made. It has been assumed that the Paleozoic rocks in normal sequence at Western Peak, Roberts Creek Mountain, and Lone Mountain represent limbs of folds (Pl. 4); only by this type of deformation is it possible to account for the high angles of dip shown by these tilted blocks. The same assumption is made for the Carboniferous and Permian rocks in the Sulphur Spring Mountains as well as Devonian strata to the east. Folding would also partly explain the distribution of the Middle Devonian Nevada formation, which appears as windows through the upper thrust plate (Pl. 4). During or following movements responsible for these folds three different kinds of faults developed. Several transverse faults have caused offsets in the Paleozoic rocks between Western Peak and Roberts Creek Mountain (Pl. 4). Associated with these is a minor low-angle thrust separating the dolomites of Cooper Peak from underlying Paleozoics. This subordinate thrust plate is apparently of local origin, made up of brecciated and dolomitized lower Nevada, probably Lone Mountain, and perhaps older rocks. Displacement of the upper plate from west to east is suggested by westerly dip of the thrust plane and by drag and overturning in Nevada limestone below the thrust sole at Cooper Peak. Apparently after the Cooper Peak thrust formed, normal faulting down-dropped Nevada limestone to the west and southwest of Roberts Creek Mountain (Pl. 4).

Following initial folding and local faulting the main Roberts Mountains upper thrust plate appears to have moved eastward, carried along by competent beds now eroded away, while the somewhat incompetent Vinini acted as the sole of the thrust sheet. A major fault of this nature might be expected to cut folds of the underlying rocks, producing a relatively regular and uniform thrust surface. On the contrary, small isolated windows of Nevada limestone (Pl. 3, fig. 3) now projecting through the eroded thrust plate indicate that the thrust surface is quite irregular. Most of the Nevada windows appear too small and too steep-walled to be explained wholly by later warping or faulting, but give the impression of having been prominences or monadnocks rising above a pre-thrusting topographic surface of relatively low relief. Under present conditions of erosion the Nevada limestone is quite resistant as compared to associated rocks.

While no positive evidence for movement of the Vinini plate over an erosion surface is forthcoming, the highly irregular features of the thrust surface described above might be expected under these conditions. In this connection mention of reported surface thrusting in Nevada and Utah may be significant. In the Muddy Mountains Longwell (1926, p. 570) found coarse conglomerate at the base of a thrust plate with pebbles and cobbles derived from the thrust plate itself filling depressions in the underlying Jurassic sandstone. Many of the pebbles have been sliced repeatedly by shearing, and nearly all were crushed. Muller and Ferguson (1939, p. 1618) describe similar features in the Hawthorne and Tonopah quadrangles, Nevada, while Nolan (1935, p. 57) suggested that a thrust in the Gold Hill district, Utah, moved over an erosion surface.

After thrusting considerable erosion took place, so that no measure of the original thrust plate thickness can now be given. A cycle of igneous activity ensued, followed by normal faulting which resulted in dislocation of the thrust sheet and its local cover of volcanics (Pl. 4). Warping or folding of the thrust plate preceding later normal faulting cannot be wholly excluded. However, apparent lack of coincidence of the many irregularities on the thrust surface with structures in underlying rock suggests that subsequent warping may not be a major factor.

Dating of the structural events is unsatisfactory in absence of adequate post-Paleozoic stratigraphic evidence within the limits of this study. In the Great Basin and surrounding regions where thrusts are becoming widely recognized the geographic limits of Nevadan (late Jurassic) and Laramide (late Cretaceous) influence are not fully understood; possibility of both earlier and later compressional diastrophism must also be entertained.

Within the area under consideration the youngest rocks involved in the major folding are undifferentiated Carboniferous and Permian. Ordovician rocks make up the entire upper thrust plate, and the youngest strata definitely covered by the thrust plate are of Devils Gate (Upper Devonian) age. Probably the undifferentiated Carboniferous and Permian in the Sulphur Spring Mountains were covered by the thrust, but at present all outcrops of these are separated from the thrust sheet by normal faults. Hence the only positive evidence is that the folding is later than rocks regarded as Permian and may thus be post-Paleozoic. The major thrusting followed folding, with perhaps no great time interval separating the two events.

Further speculations are of necessity based on analogy with thrusting in other areas. Nolan (personal communication) in the near-by Eureka mining district finds fresh-water Cretaceous involved in eastward thrust movement, thus providing a possible lower limit for corresponding events in the surrounding territory.[4] Assuming a middle or upper Tertiary age for the post-thrusting volcanics of the Roberts Mountains, it is therefore suggested that the major thrusting in question took place either in later Cretaceous or early Tertiary time.

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[4] According to Nolan (personal communication) the fresh-water Cretaceous of the Eureka district may have been deposited during later phases of the thrusting

POST-THRUSTING IGNEOUS ROCKS

GENERAL STATEMENT

After the widespread thrusting of the Vinini formation and the subsequent erosion, a cycle of igneous activity began. In Whistler Mountain, a stock of alaskite and sill offshoots have intruded the Vinini as well as later Paleozoic formations. No evidence is available to place these intrusives in the known igneous sequence which began with the intrusion of rhyolite porphyry plugs. Contemporaneous with or following the intrusion of the plugs, rhyolitic tuffs and breccias containing porphyry fragments filled canyons cut into the thrust plate. These fragmental deposits were covered by sheets of andesitic lava. Somewhat similar andesite dikes are abundant in the basement, particularly in the Paleozoic limestones, and these possibly served as feeders for the lava flows. In the northeastern corner of the Roberts Mountains, quartz latite plugs and lava flows piled up an impressive volcanic field. This activity was followed by erosion, for quartz latite fragments in clastic deposits north of Garden Valley are overlain by olivine basalt flows. A late andesite flow also covers the quartz latite in one area.

ALASKITE STOCK AND ALASKITE PORPHYRY SILLS

The largest intrusive body in the Roberts Mountains area exposed on Whistler Mountain is a muscovite alaskite stock 6 miles long and 2 miles wide. The elongation is roughly parallel to the strikes of the shales and cherts of the nearby Vinini formation. The isoclinal folding in the Vinini formation may indicate that the alaskite stock made way for itself largely by pushing and thrusting aside the country rock. No frozen inclusions of country rock were found which might favor stoping. Metamorphic effects on the Vinini formation are slight. The shales appear more indurated, but no recrystallization was recognized in thin section even in specimens collected near the contact. Gray limestones also show no indication of recrystallization.

The alaskite is rather fine-grained and holocrystalline; quartz, feldspar, and muscovite can be recognized with the lens. Some sporadic biotite is present. The petrographic details are discussed in a later section. The alaskite resembles aplite, but thin sections reveal a hypidiomorphic texture rather than the typical allotriomorphic texture of aplites, and for this reason the rock has been termed an alaskite. Directly south of the main peak, the stock contains locally spherical and ellipsoidal tourmaline nodules as much as 2 inches in diameter. Most of these nodules consist of tourmaline and quartz, thin sections revealing that the tourmaline has replaced the feldspar and mica, but in a few the center is composed of random fibers of pure tourmaline in which even the quartz has been replaced.

At the south end of the stock, sills of fine-grained alaskite and muscovite alaskite porphyry parallel the Paleozoic rocks immediately north of Devils Gate. These sills have intruded the Devils Gate limestones (Upper Devonian) and a sequence of quartzites and black spheroidal shales which may represent the Diamond Peak Carboniferous. No metamorphism is recognizable, and even the shales show little indication of induration. At least four and possibly more sills are present, but owing to complications of later faulting the exact sequence was not worked out. The sills range from 20 to 50 feet in thickness, and they possess a rude columnar jointing as well as platy jointing parallel to the sill contacts.

The rocks composing the sills are light gray to white, and some of the sills are aphyric with a fine crystalline texture. Other sills are porphyritic with scattered smoky quartz, dull feldspar, and brilliant muscovite phenocrysts imbedded in a finely crystalline groundmass. In all the sills, iron and manganese oxide stains permeate the rocks and coat the joints so that it is impossible to obtain specimens free from this secondary coating.

Since these sills have the same mineral assemblage and are close to the alaskite stock, clearly they represent offshoots from the main stock and must have been intruded at or about the same time. Unfortunately, all that can be stated concerning the age of intrusion is that it is post-thrusting. If the thrusting took place in the late Cretaceous or early Tertiary, the alaskite intrusions must be Tertiary, but no statement can be based on direct evidence The general north-northwest alignment of the several intrusives—alaskite stock, porphyry plugs, and quartz latite domes—possibly suggests a structural control and general contemporaneity. A comparable stock of quartz monzonite has been reported by Nolan (1935, p. 43-48) in the Gold Hill district as Eocene or Oligocene, so that the alaskite may be early Tertiary. But no evidence contradicts middle Tertiary age.

RHYOLITE PORPHYRY PLUGS AND DIKES

Three large and three small plugs of porphyry have been mapped in the present study In addition narrow dikes of similar rock were found chiefly in the region of Garden Pass, but the map scale prohibits their representation. The most southerly of these plugs comprises Mount Hope (Pl. 2, fig. 3). In addition to the main plug, small dike offshoots occur at several places on the lower slopes of Mount Hope. Fresh samples of the Mount Hope intrusive are difficult to obtain for the feldspar phenocrysts are dull and chalky, and the rock has a gray dull luster. Occasional fresh samples are brilliant white with numerous scattered round and corroded smoky quartz phenocrysts associated with pale-cream colored sanidine phenocrysts. Plagioclase and mafic minerals are absent. The groundmass is very finely crystalline. The chemical analysis indicates that this plug is composed of potash rhyolite porphyry. It was reported to the writers that the Mount Hope porphyry plug has been mineralized with sphalerite and galena, and abandoned mine workings are located at the southeastern margin of the plug. Undoubtedly this mineralization is responsible for the altered appearance of most of the intrusive. The dikes in the Garden Pass region are also potash rhyolite porphyry.

The plugs to the northwest of Mount Hope are similar except that the quartz phenocrysts are usually more smoky, and plagioclase is an important constituent along with sanidine. Also scattered small crystals of biotite and reddish garnet appear against the grayish-white finely crystalline groundmass. The unusual appearance of this garnet is characteristic of these intrusives and is of some importance in relating these intrusives to the later igneous history, as will be emphasized later. Zirkel (1876, p. 195) recognized this garnet in a specimen from this region in connection with the Fortieth Parallel studies. The two largest plugs are 2 miles northwest of Mount Hope, and, in addition to intruding the Vinini formation, one has penetrated a window of Nevada limestone. A small plug occurs on the south slope of Vinini Creek Canyon also northwest of Mount Hope, and two are present on the eastern slope of Roberts Creek Mountain 2 and 2½ miles respectively southeast of the summit.

TUFFS AND BRECCIAS

The next episode in the geologic history is represented by deposits of rhyolite tuff and breccia which filled a canyon or canyons cut in the Roberts Mountains thrust plate. The thickest accumulation of these deposits is 4 miles south-southwest of Roberts Creek Mountain where 700 feet of tuffs and breccias appear beneath the lava cap. They thin rapidly to the north and south so that, at the northern edge of the largest lava mesa, only a few feet of breccia separates the Vinini from the lava cap. As a rule, exposures of the tuff are poor, for the overlying lava tends to slide or to break into talus covering the tuff. In the thick section of tuff, scattered exposures give a general sequence as follows: Above the Vinini formation is a thick deposit of white massive pumice lapilli tuff in which quartz, sanidine, and plagioclase crystals are prominent. Above this is a breccia of the garnet-bearing porphyry with complete absence of glass or pumice fragments. The fragments of porphyry in the breccia vary in size up to 1 foot or more in diameter. Overlying the porphyry-breccia is a thick layer of grayish coarse vitric-crystal tuff, according to the nomenclature of Wentworth and Williams (1932, p. 52). This particular tuff is well indurated, and thin sections reveal recrystallization of the glass shards indicating some incipient welding during or shortly after emplacement. The remainder of the visible section consists of pink, massive pumice-lapilli vitric-crystal tuff in which large quartz, sanidine, plagioclase, and biotite crystals are distinct. The crystals are of the same size and character as those in the garnet-bearing porphyry plugs; however, no garnet was observed in the tuff. On the north side of the large lava mesa, scattered fragments of garnet-bearing porphyry indicate the presence of porphyry-breccia directly under the lava, but the thickness could not be determined.

In the belt of tuff 2 to 3 miles north and northeast of Roberts Creek Ranch, 200 feet of tuff is present with the base not exposed. The exposures are very poor, but several reveal that the tuff is partly stratified and sorted, indicating water transportation. The southern margin of the andesite cap rests on the Vinini formation. North of Vinini Creek, 20 feet of fine water-laid vitric tuff rests on the Vinini formation and is in turn covered by the andesite lava cap. Two miles west and southwest of Roberts Creek Mountain, a block of andesite has been downfaulted, and tuff is poorly exposed at the margins of the andesite. At the southern end of the lava cap, 40 feet of white vitric-crystal tuff similar to that described above is exposed beneath a thin layer of porphyry-breccia. Indeed, around the margins of several of the andesite caps, scattered angular fragments of the garnet-bearing porphyry indicate the widespread distribution of this breccia prior to the outpouring of the andesite flows. Because of the purity of these breccias, undoubtedly the plugs were intruded at the same time as the tuff and breccia series accumulated, and porphyry intrusives must have reached the surface as domes. Avalanching and rapid erosion of these domes would supply the angular fragments of porphyry in great abundance and purity needed to form these breccia sheets. If the porphyry had been uncovered by erosion, fragments of the cover rock would certainly be present in the breccia. Therefore it is suggested that the garnet-bearing plugs were protruded to the surface as domes contemporaneously with explosive eruptions of rhyolitic ejecta. Some of the ejecta was later reworked by running water, accounting for the stratified and sorted tuffs. It is quite possible that the massive pumice-lapilli vitric-crystal tuffs represent the products of pelean blasts, for otherwise the tuffs would be better sorted and stratified.

The tuff deposits, particularly the stratified and sorted tuffs, are younger than the thrusting, and fossils may be found permitting the dating of this igneous cycle. In the water-laid tuffs northeast of Roberts Creek Ranch fragmentary plant stems, leaves, and indeterminate silicified coniferous woods were found.

ANDESITE FLOWS AND DIKES

Reference has already been made to the andesite which appears as isolated lava caps (Pl. 1, fig. 1) on the eastern and southern slopes of Roberts Creek Mountain as well as in a downfaulted block west of the mountain. Undoubtedly these flows were formerly much more extensive, and only erosional remnants remain. The thickest section of andesite is 4 miles southeast of Roberts Creek Mountain, where 200 feet of lava is exposed. At least four flows are present, each flow marked by a vesicular top and massive interior. The flows range from 40 to 60 feet thick. Although there are some variations in appearance, vesicular phases being rather black and dull, these flows are usually medium gray and finely crystalline. Phenocrysts are rare. Vesicles are spotted with cristobalite crystals, and irregular cavities lined with projecting crystals of feldspar are in turn coated with small spheroidal-shaped crystals of cristobalite. In the lava cap 2½ miles southeast of Roberts Creek Mountain, only two flows remain, and elsewhere on the eastern slope of Roberts Mountain only one flow can be found. In the block southwest of Roberts Creek Mountain, two or more flows are probably present, but owing to the local landslides at the margin of the flows the number was not certainly determined.

As the rhyolitic tuffs and porphyry-breccia filled canyons cut largely in the Vinini formation, and all the andesite flows have some of this material underlying them, probably the lava closely followed the rhyolitic ejections and poured down the upper portions of these valleys. Whether or not they completely filled the canyons and spread out over the entire surface cannot be answered owing to the subsequent erosion, but the limitation of these flows to areas covered with rhyolite tuffs indicates that the lava did not completely bury the region.

On the western slopes of Roberts Creek Mountain a number of andesite dikes cut the Paleozoic limestones. They are of variable width ranging up to 40 feet and are essentially vertical. Five similar dikes were noted north and east of Cooper Peak, one cutting the Vinini formation. With this one exception, all dikes noted were in limestones, but considering the methods of mapping possibly they were missed in the Vinini because of the appreciable mantle burying the dike outcrops. Even in the limestones, the dikes are appreciably weathered and usually outcrop in gaps in the ridges. As a result of weathering, the andesite is deeply impregnated with iron oxide so that fresh samples are difficult to obtain. The andesite dikes are slightly more crystalline than the andesite in the lava flows and contain no cristobalite in cavities, but otherwise they have essentially the same texture and mineral composition. For these reasons they are probably essentially contemporaneous, and the dikes possibly represent feeding dikes for the flows.

QUARTZ LATITE FLOWS

The northeastern corner of the Roberts Mountains is covered by thick flows of quartz latite, and several patches lie to the north of Vinini Creek. Also two isolated domes appear near Roberts Creek Mountain, the larger less than 1 mile to the north and the smaller 1½ miles to the southeast. East of Dry Creek, near the western margin of the quartz latite, 75 feet of massive greenish-gray vitric-crystal quartz latite tuff underlies a friable gray quartz latite flow. Apparently the first eruptions of quartz latite, at least locally, were explosive. In the reconnaissance studies, no other exposures of tuff were located, and, if it is widely distributed, it has been covered by the extensive quartz latite flows. Doubtless detailed mapping will disclose other exposures.

The bulk of the quartz latite is pink to reddish owing to the hematite in the groundmass. The groundmass is usually aphanitic, and flow banding may be present Locally some of the quartz latite is dark gray with a perlitic glassy groundmass. Everywhere it is in contact with the reddish quartz latite, the perlitic variety is the older, suggesting that it was the first quartz latite lava erupted.

Apparently many vents served as sources for the quartz latite lava. Many of the summit areas in the lava field have the characters of domes only slightly modified by erosion, such as the domelike shape and flow banding which is parallel to the margin of the dome and dips inward on all sides. In addition, one dome contains an appreciable amount of quartz latite breccia, apparently formed as the solid quartz latite was forced upward. These characters are common to other domes (Williams, 1932, p. 144-5). Local areas of quartz latite in the domes have been bleached white as if by solfataric action, also a common feature among recent domes. Much quartz latite represents thick viscous flows that may have poured out of these vents prior to the dome formation; in part the domes represent the last upward protrusion of lava too viscous to flow. A comparable case has been described in a recent flow elsewhere (Anderson, 1933, p. 491). Some of the domes may represent crater infillings, after the eruption of the lava flows.

Two miles northwest of Oak, a thin layer of porphyry breccia is assumed because of the scattered fragments in the mantle. Directly above these fragments, a flow of andesite similar to that found to the south and southwest is directly covered by the quartz latite, thus giving positive evidence of the sequence: tuff, andesite, quartz latite.

POST-QUARTZ LATITE VOLCANIC ACTIVITY

Near the northeastern margin of the Roberts Mountains, a single flow of dark-gray pyroxene andesite makes up the summit of a small hill. It is surrounded and underlain by quartz latite, showing that it is of post-quartz latite age. In the field, this late andesite somewhat resembles the andesites of pre-quartz latite age to the south, but in thin sections it is entirely different. Obviously it is an erosional remnant, but the source vent and its original distribution are unknown. Since the quartz latite apparently did not undergo much erosion, this late andesite may have been a flow of local source and limited distribution.

The southern end of a small north-south fault block ends at the northern end of Garden Valley, and in the eroded fault scarp, which bounds the western edge of the block, poorly stratified tuff contains angular fragments of quartz latite. Olivine basalt overlies the tuff beds and makes up the eastern back slope of the fault block. The quartz latite fragments in the tuff indicate that the olivine basalt is post-quartz latite.

OTHER VOLCANIC ROCKS

At the western margin of the Roberts Mountains are two areas of volcanic rocks, but their relationship to the other volcanic rocks is unknown. The more northerly outcrops are reddish porphyritic hornblende andesite, totally different from the andesites on the eastern portion of the mountain. It resembles andesites in the Simpson Park Mountains, west of Roberts Mountains.

Two small hills of rhyolite rise above the alluvium west of the mouth of Rutabaga Creek. Only about 40 feet of rhyolite remains. At the base of the northwestern hill, a few feet of hard well-cemented rhyolite breccia is exposed beneath massive rhyolite, suggesting that the rhyolite represents a flow rather than an intrusive body. Phenocrysts of smoky quartz, plagioclase, sanidine, and biotite recall the garnet-bearing porphyry plugs, although no garnet was observed in these rocks. However, some of the tuff beneath the older andesite flows has similar crystals, and perhaps this rhyolite at the southwest margin of the Roberts Mountains is related to the rhyolite tuff series.

AGE OF THE IGNEOUS ROCKS

No evidence is available to date the igneous rocks except that they are later than the thrusting, and if this event is of late Cretaceous or early

Tertiary age the igneous rocks are necessarily Tertiary. Indeed most geologists would suggest a Tertiary age solely on the basis of comparable rocks elsewhere in Nevada that have been dated by means of fossiliferous continental beds. However, since lithologic similarity is dangerous for correlation of lavas and intrusive rocks from one district to another, the precise age of the igneous rocks in the Roberts Mountains is an open question. However, in summary the following sequence should be emphasized: (1) Rhyolite tuffs and breccias are contemporaneous with the protrusion of porphyry plugs and (2) intrusion of dikes followed by andesite flows, (3) quartz latite, and (4) later pyroxene andesite and olivine basalt. The alaskite stock and its sill offshoots, and the hornblende andesite and rhyolite on the western flank of Roberts Mountains, cannot be placed accurately in the sequence.

POSTVOLCANIC FAULTING

The andesite flows in the Roberts Mountains give evidence of the character and magnitude of the faulting which followed their eruption. Southwest of Roberts Creek Mountain, a narrow block of lava and underlying Vinini has been downfaulted against the Vinini formation on the west and Paleozoic sequence on the east. Projecting the lava westward from the exposures on the south-eastern flank of Roberts Creek Mountain to the south end of the narrow block, a net displacement on intervening faults is roughly 1500 feet. The andesite surface rises to the northwest some 800 feet in elevation, and the net displacement is probably correspondingly less. Apparently the eastern fault of this block followed a pre-existing fault causing the downdropping of the Nevada limestone prior to thrusting. Transverse faults terminate this central fault block at both the northern and southern ends. At least two other blocks have been downdropped south-southeast of Roberts Creek Mountain, and here also the andesite flows indicate the displacement. At least a 1000-foot displacement is represented in the small block 4 miles north of Roberts Creek Ranch, and a 400-foot displacement separates this block from the adjoining southwest block.

The thrust plate has been downfaulted against the undifferentiated Carboniferous rocks at the west margin of the Sulphur Spring Mountains, and probably the normal faults between the Vinini and Carboniferous rocks developed at the same time as the normal faulting of the thrust plate to the west.

OUTLINE OF ROBERTS MOUNTAINS BLOCK

Geologists today commonly assume that many of the present mountain ranges in Nevada represent fault blocks with variations, their outline determined largely by boundary normal faults. Apparently the age of the faulting varies, for with some ranges slightly eroded fault scarps are characteristic, and with others well-developed pediments indicate considerable recession of the original fault scarp. The Roberts Mountains and neighboring Sulphur Spring Mountains with their extension to Whistler Mountain do not have a simple outline that one can easily relate to boundary faults, so that the presence of faults where steep slopes rise above alluvium is sometimes doubtful. The eastern margin of the Sulphur Spring Mountains, although somewhat irregular, is in general sharply set off from Diamond Valley, and on physiographic evidence presumably the western margin of Diamond Valley is bounded by a normal fault. Hot springs along this margin are further suggestive evidence. However, the outline of the range is irregular, particularly in the southern extension. Where the Vinini formation makes up the southern extension, a wide pediment is present, for occasional gullies show that the alluvium represents only a rather thin veneer resting on a surface cut on the Vinini. Where the alaskite stock appears, the pediment is much narrower. It is suggested that as the rocks of the Vinini formation are intensely jointed, they are rather susceptible to mechanical disintegration in the desert environment so that the original fault scarp has been rapidly eroded back with the simultaneous development of a pediment. In the main, the alaskite has joints much more widely spaced and is slowly attacked by weathering so that the pediment to the east of the alaskite stock is developed solely in the Vinini. This relation between width of pediment and bedrock character has been emphasized by Gilluly (1937, p. 329). A suggestion of renewed faulting with uplift of the pediment surface is faint 1 mile east of the alaskite. If the small scarp, 20 feet high and somewhat eroded, gives the location of the original boundary fault, the pediment is about 1 mile wide east of the alaskite and 2 miles wide east of the ridge of Vinini formation. Farther north, the Nevada limestones apparently have receded but little except locally.

At the western margin of the Roberts Mountains is a well-developed pediment particularly from Roberts Creek Ranch to Jackass Creek. This pediment is broken by a north-south fault which passes through the Three Bar Ranch. The evidence for faulting is entirely physiographic; the pediment surface, veneered by alluvium, is cut off to the west by a north-south scarp 75 to 100 feet high. Possibly this scarp represents the location of an original boundary fault.

The northern boundary of the Roberts Mountains is rather straight and steep with exception of the northwestern area where the Vinini formation is exposed, and here a well-developed pediment is present. Physiographic evidence suggests that the northern boundary is a normal fault, but, owing to the character of the rock, little recession of the scarp has taken place.

Southeast of Roberts Creek Ranch a ridge capped by lava apparently represents a block bounded by faults on both sides—faults that are extensions of the faults in the Roberts Mountains that bound downdropped blocks. For this reason, it is possible that the internal faults in the range bordering downdropped blocks formed at the same time as the boundary faults outlining the Roberts Mountains and Sulphur Spring Mountains.

That there has been some recent faulting east of Roberts Mountains is indicated by a scarp that cuts the alluvium south of Oak in Garden Valley. The scarp is over 100 feet high 3 miles south of Oak and can be traced for about 4 miles. The alluvium appears to have been washed down the eastern slopes of Roberts Mountains and carried to the north by Henderson Creek. Garden Valley probably has a structural origin, but later detailed work may modify this suggestion.

The origin of isolated Lone Mountain, surrounded by alluvium, presents an interesting problem. It may represent a block bounded by normal faults, but no supporting evidence was found. More likely it was an area of some relief on a downfaulted block in the formation of Kobeh Valley. The base of Lone Mountain is in part composed of white diatomaceous siltstones resembling lacustrine sediments. These beds have been dissected by numerous gullies and are veneered with fanglomerate washed down from Lone Mountain. A Pleistocene lake in the Kobeh Valley is indicated by Meinzer (1917, Fig 1) as well as one in Diamond Valley to the east. As an intermittent stream now drains Kobeh Valley into Diamond Valley through Devils Gate, probably the two Pleistocene lakes were connected, and part of the deepening of Devils Gate probably was the result of overflow. With the lowering of the outlet and disappearance of the lake, the sediments deposited in the Pleistocene lake would be eroded to their present form. The blanket of Pleistocene sediments around Lone Mountain effectively masks the base so that its origin must remain doubtful.

PETROGRAPHY OF IGNEOUS ROCKS

GENERAL STATEMENT

Details concerning the petrography of the post-thrusting igneous rocks are considered separately in order to emphasize the relations and differences among the several groups. The descriptions of the Ordovician igneous rocks, given earlier, are not here elaborated as they differ markedly from the later rocks.

ALASKITE AND ALASKITE PORPHYRY

The alaskite is fine-grained hypidiomorphic granular. The feldspar consists of sodic oligoclase and orthoclase, both somewhat clouded. The oligoclase is in small euhedral to subhedral crystals (.15 mm) with faint to sharp albite twinning. The orthoclase is usually anhedral with a few subhedral grains. The fine grain and clouded appearance make it difficult to estimate closely the ratio of plagioclase to orthoclase; the oligoclase appears to be in excess. The feldspars together account for about 66 per cent of the rock. Quartz is commonly in slightly larger grains (.2 mm), molded about the smaller oligoclase crystals. A few quartz grains poikilitically enclose oligoclase. The quartz content is approximately 32 per cent, as measured by a micrometer stage, and the normative quartz is about the same (Table 2, no. 4). The remaining 2 per cent consists largely of anhedral muscovite up to .3 mm long, molded about the other minerals, in places completely enclosing both the quartz and feldspar. Biotite, partly altered to chlorite, is rare in some specimens, either intergrown with muscovite or in minute anhedral crystals. Zircon and magnetite are minor accessory minerals. The border phase of the alaskite is very fine-grained (0.07 to .08 mm.) with the above minerals in the same proportions; no tendency toward a porphyritic texture is visible.

The tourmaline nodules mentioned earlier consist largely of tourmaline and quartz. In a few of the larger nodules, the core is composed of randomly oriented acicular crystals of black tourmaline, with an outer zone a quarter of an inch thick consisting of quartz and tourmaline. In some nodules minor amounts of feldspar may be observed in thin section. The edges of the nodules are irregular but sharp. Doubtless the tourmaline has developed largely by the replacement of the feldspar, and such remnants of feldspar as are still preserved are usually oligoclase, indicating that the orthoclase is more readily replaced. Around one nodule, muscovite interstitial to the quartz is abundant and presumably developed by replacement of feldspar. Under low magnification the quartz grains in the nodules are obviously larger and more numerous than in the enclosing alaskite. This is in keeping with the suggestion of Edwards (1936) that orthoclase plus boron yields tourmaline and quartz. Edwards in a study of comparable nodules was able to recognize later deposits of quartz. None were observed in this study, but the larger size of the quartz grains indicates addition.

The tourmaline is usually in irregular short crystals interstitial to the quartz; distinct prisms were rarely observed. In a few examples neighboring grains extinguished simultaneously, indicating a common orientation and presumably representing a poikiloblastic texture. The tourmaline is pleochroic from medium blue to pale gray. The indices of refraction are $\varepsilon = 1.638 \pm .003$, $\omega = 1.670 \pm .005$. These indices, birefringence, and color indicate the variety schorlite, suggesting that iron as well as boron must have been added in order to develop the tourmaline.

No observations contradict Edwards' suggestion (1936) that these nodules develop by pneumatolytic replacement of feldspar *in situ* after the intrusion of the host rock. Tilley (1919) has pictured the development of bubbles of mineralizers which rose in the viscous magma until they reached the crystal mesh of quartz and feldspar and attached themselves to these crystals, attacking the feldspars and converting them into tourmaline at the same time that they deposited the excess silica as secondary quartz.

The lowermost sill in Devils Gate is aphyric and in thin section resembles the chilled border phase of the main alaskite stock except that the grain size is slightly larger, and a little more muscovite is present. The remainder of the sills are porphyritic with numerous sodic oligoclase phenocrysts and random quartz (1 to 2 mm.) and muscovite phenocrysts (1 to 1.5 mm.) set in a microgranular aggregate of quartz, feldspar, and muscovite, the grain size averaging about .05 mm. in dimension. Owing to the small size of the crystals and clouding of the feldspar it is impossible to differentiate between the orthoclase and plagioclase in the groundmass.

RHYOLITE PORPHYRY

The rock from Mount Hope has a distinct porphyritic texture with quartz and sanidine phenocrysts ranging up to 5 mm, the quartz phenocrysts in slight excess of the sanidine. The groundmass consists entirely of microcrystalline quartz and feldspar, and judging from the chemical analysis (Table 2, no. 6) the groundmass feldspar is also sanidine. The feldspar is in part micropoikilitic and contains quartz. Most of the rock from the Mount Hope plug is sericitized, the feldspar phenocrysts and groundmass altered and some secondary quartz deposited in disconnected veinlets. A sample with a small amount of sericite was selected for analysis (Table 2, no. 6). It is remarkable in the exceedingly high potash and low soda content. This rock should properly be termed a potash rhyolite porphyry.

In the garnet-bearing rhyolite porphyry, the smoky quartz phenocrysts average about 3 mm, and many show pyramidal faces. The feldspar phenocrysts are slightly smaller and include calcic oligoclase and sanidine in approximately equal amounts. Under the microscope the feldspar phenocrysts seem to be clustered. Biotite is rare and in some specimens is of the greenish variety associated with chlorite; normally it shows intense pleochroism from pale yellow to very dark brown.

Reddish garnet crystals, ranging up to 2 mm, are scattered throughout the rock but represent less than 1 per cent of the specimen. In thin section some of the garnet is observed to be associated with quartz or feldspar phenocrysts, and some is isolated. The garnet is isotropic and distinctly pink in thin section. From its association and color, it probably belongs to the almandite-spessartite series, Pabst (1938) having described garnets of this series from lithophysae in rhyolites from Ely. The groundmass of the porphyry contains scattered oligoclase microlites separated by a micropoikilitic aggregate of quartz and untwinned feldspar, accompanied by some iron ores. The similarity in composition of the garnet-bearing porphyry to the alaskite (Table 2, nos. 4 and 5) is very striking. Possibly this indicates a close genetic relationship with different modes of formation. Also the contrast to the potash rhyolite porphyry of Mount Hope should be noted (Table 2, nos. 5, 6); the garnet-bearing porphyry contains appreciable soda and much more lime. The garnet-bearing porphyry appears to be a normal rhyolite porphyry.

The rhyolite flow near the mouth of Rutabaga Creek is similar to the above description of the garnet-bearing rhyolite porphyry except that the biotite has been largely replaced by granular magnetite, and the latter is more abundant in the groundmass. Garnet is absent.

QUARTZ LATITE

The average quartz latite is gray to pink with conspicuous white plagioclase phenocrysts 3 to 6 mm long. Associated phenocrysts are round glassy quartz up to 3 mm in diameter and dull black biotite up to 2 mm in diameter. In thin section, the plagioclase exhibits sharp albite twinning and faint zoning. Some glass inclusions may be present. Refractive indices indicate that the feldspar is largely andesine to calcic andesine, but in one rock the phenocrysts are sodic labradorite. The quartz is usually corroded and subordinate to the feldspar. In some samples of quartz latite, quartz is rare. Most of the biotite is replaced by granular magnetite, but a few shreds remain to indicate the original mineral. In a few thin sections, unaltered biotite is deeply pleochroic from pale yellow to deep brown. In some samples, small reddish-brown basaltic hornblende is present, rarely rimmed by magnetite. Accessory minerals are numerous apatite prisms and some zircon as well as magnetite. The groundmass ranges from isotropic to microcrystalline. Isotropic areas are local and never comprise an entire thin section; the average groundmass is cryptocrystalline with a "patchy" appearance suggestive of devitrification. In some thin sections the groundmass is composed of a mat of minute spherulites, .03 mm in diameter. Flow banding can be observed in some rocks. It is accentuated by red iron oxide which is irregularly scattered throughout most of the quartz latites and accounts for their general pink color. In one specimen of pronounced flow banding, spherulites, 2 and 3 mm in diameter, are arranged parallel to the banding; these spherulites only feebly polarize light. In some sections, the quartz latite may be partly granulated, no doubt owing to protrusion in a highly viscous to solid state.

TABLE 2.—*Chemical analyses and norms*

	1	2	3	4	5	6
			ANALYSES			
SiO_2	55.56	57.66	69.35	74.15	74.97	76.15
Al_2O_3	15.15	13.59	15.22	14.96	14.12	12.75
Fe_2O_3	3.18	4.30	2.50	.17	.44	.00
FeO	6.34	6.26	38	64	.53	09
MgO	4.33	2.79	51	.15	09	.07
CaO	8.15	5.86	2.40	69	.83	04
Na_2O	2.97	2.82	3.25	3.72	3.77	.50
K_2O	1.75	2.28	4.48	4.36	4.49	10.38
H_2O+	59	1.10	.67	.70	.56	.09
H_2O-	22	93	66	06	.17	00
CO_2	..	45	....	....	....	....
TiO_2	1.29	1.53	.40	03	02	.05
P_2O_5	41	.51	15	10	02	.03
MnO	16	15	.02	07	.08	Tr
	100.10	100.23	99.99	99.80	100.09	100.15
			NORMS			
Q	8.58	15.22	27.21	33.62	33.61	31.38
Or	10.56	17.24	26.69	26.13	26.69	60.78
Ab	25.15	23.58	27.77	31.44	31.96	2.22
An	22.80	15.85	11.12	2.78	3.89	.....
C	..	...	.82	3.06	1.53	..
Di	12.46	5.69	......	...	...	...
Hy	11.88	10.02	1.30	1.32	73	20
Il	2.43	2.89	.76	....	...	15
Mt	4.64	6.26	.23	23	.70	
Hm	...	....	2.40	...	..	.
Ap	1.01	1.34	34	34	..	
Cc	.	2.20	...	...		

1. Andesite from large mesa 4½ miles northwest of Mount Hope
2. Andesite dike from 9000 feet elevation, west slope of Roberts Creek Mountain
3. Quartz latite from summit of hill in southeast corner of quartz latite area
4. Alaskite from near top of Whistler Mountain.
5. Rhyolite porphyry from plug north of Mount Hope.
6. Potash rhyolite porphyry from Mount Hope.

All analyses by G. Kahan.

The dark-gray perlitic quartz latite contains the same phenocrysts, including the basaltic hornblende, as the average pink quartz latite but in addition has hypersthene, with pronounced pleochroism. The perlitic groundmass is charged with numerous microlites and crystallites separated by clear glass with a refractive index of 1.501 ± .003.

A microscopical examination indicates a normal dacite, but the chemical analysis (Table 2, no. 3) shows that the potash content (4.48 per cent) is far too high for such classification, and it is more appropriately called a quartz latite. As with many latites, the potash must be present largely in the groundmass, with the exception of a small amount in the biotite.

ANDESITE

The older andesite flows that antedate the quartz latite are composed almost exclusively of labradorite, augite, and magnetite. The texture is aphyric; visible phenocrysts are rare and consist almost entirely of plagioclase, usually calcic labradorite, 1 to 2 mm long. Except for the surface vesicular phases, the andesite has an intergranular to intersertal texture, and the labradorite crystals range from An_{50} to An_{60} and are three times as long as the greenish-gray augite granules. The labradorite grain size varies from .1 mm long in one extreme to an average of .3 mm in another. In one slide, an augite microphenocryst has a hypersthene core; normally this latter mineral is absent. A small amount of olivine was noted in one sample. The magnetite may appear as distinct crystals or as dust in the glass if the texture is intersertal. If all the magnetite is in large crystals, the rock is light gray. Apatite is a common accessory. Cristobalite is present in practically all samples of the andesite, either as spheroidal .2 to .3 mm in diameter lining vesicles or more commonly coating irregular-shaped cavities into which the feldspar projects. The labradorite in these cases may have a very narrow outer rim of albite which is in optical continuity with the labradorite. If the texture is intersertal, the interstitial glass varies from black opaque dusted with magnetite to pale gray shot through with long slender needles too minute to polarize light. In one slide, some of the interstitial glass is reddish, with some birefringent fibers suggestive of chlorophaeite and its alteration products (Peacock and Fuller, 1928, p. 369). The surface vesicular phase is always black in hand specimens and thin section, and the labradorite and augite are completely separated by a black opaque mesostasis, forming a hyalo-ophitic texture.

The classification of the older andesite presents a problem familiar to all petrographers, for it is not a typical andesite in that the modal feldspar is labradorite. On the other hand it is not a typical basalt in that the silica content is high (Table 2, no. 1), raised possibly 1 per cent by the cristobalite. However, the normative plagioclase is andesine, and the normative salic minerals total 67 per cent. Obviously this rock is on the border line between andesite and basalt, and some petrographers will prefer one term and some the other. Without any strong convictions, the rock is here called an andesite.

The dikes are similar to the flows, except that the dikes have no cavities and no cristobalite. Hypersthene is present in small amount in most of the thin sections, and the magnetite always appears as definite crystals. The texture is intersertal to diabasic, and the grain size varies greatly from dike to dike. In the coarsest texture observed, the feldspar crystals average .5 mm long and, in the finest, .15 mm. Where the texture is intersertal, the interstitial material may be light-colored glass riddled with slender prisms of apatite or consist of a grayish cryptocrystalline aggregate. In some dikes the interstitial material may be orange, opaque, with an index below balsam, and containing some birefringent fibers that suggest chlorophaeite and its alteration products. In several dikes, the light-colored base interstitial to the augite and labradorite weakly polarizes light and is arranged in radiating bundles. The index of refraction is below balsam, indicating that this material may be an alkali feldspar. Quartz was observed in small irregular grains in this aggregate which suggests that the final liquid crystallized to quartz and alkalic feldspar. Apatite is common in all these interstitial areas. A holocrystalline diabasic texture is not common and usually embraces only a part of a thin section. Grains and aggregates of calcite are common in all the dikes studied, and, in a few, rounded quartz grains associated with calcite indicate that some of the quartz may be xenolithic. No reaction rims were noted around either the calcite or larger quartz crystals. The freshness of the associated pyroxene and feldspar indicates that the calcite represents xenocrysts from the wall rock.

Even if the dikes represent feeders for the early andesite flows, the composition is different (Table 2, nos. 1, 2), for the analyzed dike rock is higher in silica and potash and lower in magnesia and lime, placing it definitely among the andesites.

The andesite resting on the quartz latite in the northeast corner of the Roberts Mountains is dark gray with scattered andesine and numerous small augite phenocrysts. Thin sections rarely include the plagioclase phenocrysts, but the pale-gray augite phenocrysts are numerous. The groundmass is pilotaxitic, with andesine and augite microlites dusted with magnetite.

The hornblende andesite from the west side of the Roberts Mountains is conspicuously porphyritic with large phenocrysts of andesine-labradorite and basaltic hornblende, the latter rimmed with granular magnetite. The groundmass is pilotaxitic stained with red iron oxide. Stubby faintly pleochroic yellow prisms of apatite are accessory.

OLIVINE BASALT

Olivine basalt appears in only one area, north of Garden Pass, and in the massive interior of the flow this lava is holocrystalline with an ophitic to subophitic texture. About 5 per cent consists of olivine crystals, partly altered to iddingsite, and associated with labradorite tablets .3 mm long separated from or enclosed by purplish-gray, nonpleochroic augite. Distinct crystals of magnetite and slender prisms of apatite are important accessories.

CHEMICAL CHARACTERS OF IGNEOUS ROCKS

The number of chemical analyses available from the Roberts Mountains is insufficient to permit construction of a variation diagram that has real meaning. Even when the analyses of the igneous rocks from the neighboring Eureka district are added, there is a scattering of points that allows considerable latitude in drawing the connecting curves. On the variation diagram prepared in studying the analyses, the potash curve had higher value than the soda curve where silica exceeds 60 per cent. Nolan (1935, p. 50) commented on the fact that the volcanic rocks in Utah and New Mexico have appreciable quantities of potash. In order to compare the central Nevada volcanic rocks with those of other parts of Nevada, Utah, and California the *k* ratio—that is, the molecular ratio of potash to total alkalies—has been determined from all available analyses from these areas. Since many of the available analyses record only the silica, soda, and potash content, no other ratios could be computed, and the *k* ratios have been plotted against the silica content (Fig. 3). It should be emphasized that, if the weight per cent of soda and potash are equal, the *k* ratio will equal 0.4, and any higher *k* value indicates that the weight per cent of the potash exceeds the weight per cent of soda. Figure 3 shows that most of the Utah lavas have a *k* ratio equal to or exceeding 0.4, and those below this value are not appreciably smaller. A few Utah lavas have a rather high *k* value. However, the volcanic rocks from California chosen from the Sierra Nevada and Cascade Range have a much lower *k* ratio than the Utah lavas except for the latite from Tuolumne Table Mountain, which is indicated separately in Figure 3. Also the silicic rocks from both areas have comparable *k* values. However, the intermediate and basic lavas differ considerably. The rocks from Roberts Mountains and the Eureka district, with but one exception, have *k* values close to or exceeding 0.4.

Some of the lavas from western Nevada have low *k* ratios, and others have higher values. One difficulty is that some of these lavas have been collected from mining districts such as Tonopah, Goldfield, and Virginia City, where there is always the possibility that some selective leaching of one or the other alkalies may have occurred during the process of hydrothermal alteration that accompanied metal mineralization. Terzaghi has suggested (1935, p. 378) that many potash-rich rocks are the result of alteration, and Fenner (1934; 1936) has described excellent examples of potash enrichment in the hydrothermally altered lavas of Yellowstone.

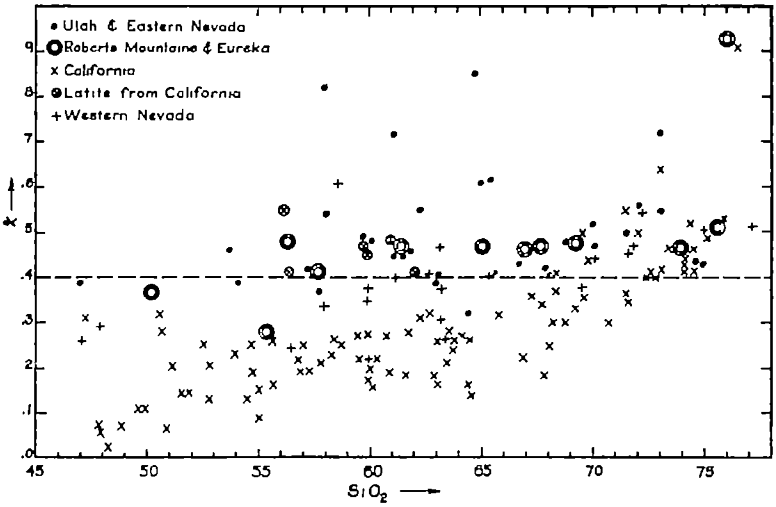

FIGURE 3—*Diagram of the k ratio plotted against SiO_2 content*

The rhyolite porphyry from Mount Hope is high in potash, and some hydrothermal alteration has taken place in the plug, although the analyzed sample was carefully selected to avoid alteration. However, as Fenner states (1934, p. 242), fresh lavas with high content of silica and potash which have not undergone hydrothermal metamorphism resulting in potash enrichment should only contain phenocrysts of quartz and orthoclase, like the Mount Hope porphyry. In the samples studied by Fenner, the potash enrichment was confined to the groundmass, and the plagioclase and pyroxene phenocrysts "remain as witnesses of the original composition, . . . incompatible with the present groundmass". So if hydrothermal alteration is responsible for the potash enrichment of the Mount Hope porphyry it differs from the Yellowstone examples in that the feldspar phenocrysts are now sanidine, and under the microscope there is no suggestion that they represent replaced plagioclase. In the other rhyolite porphyries, sanidine and plagioclase appear together. Since no definite evidence supports the suggestion that the potash enrichment followed hydrothermal alteration, it can only be considered a possibility. This same conclusion can be applied to all the rocks from the Roberts Mountains and equally well to those from Utah. Apparently the potash enrichment is regional and cannot be explained simply by hydrothermal alteration. It is not in the province of this paper to advance an explanation of the potash enrichment, but the enlargement of the Utah and New Mexico potash province to include central Nevada should be emphasized.

SUMMARY OF CONCLUSIONS

In central Nevada rocks which represent western graptolitic facies of Lower and Middle Ordovician age have been thrust eastward at a low angle over a section of sediments ranging from Cambrian to Permian. Horizontal displacement of at least 16 miles is indicated, with possibility of much greater movement. It is concluded that folding of the Paleozoic strata below the thrust preceded the major thrusting. Thrust sheet and partial cover of younger volcanic rocks have been broken by normal faults. No precise dating is assigned to the periods of folding, thrusting, or later normal faulting. The post-thrusting igneous rocks have a high potash content, unlike those in California, but are similar to the lavas of Utah and New Mexico.

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CORRELATION OF THE DEVONIAN SEDIMENTARY FORMATIONS OF NORTH AMERICA

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1. Correlation of the Devonian formations of North America (Chart 4 of series). 1788

INTRODUCTION

HISTORICAL BACKGROUND

The Devonian system became established in North America in 1847 through the writings of Verneuil, whose views were made known largely by Hall's translation. In 1851 Hall accepted the Devonian system and the European arrangement of it. Since then the system has been elaborated and expanded by the addition of the Helderberg stage. In this century a few correlation charts have been prepared (Ulrich, 1911; Schuchert, 1910; Shimer, 1934), but none considered Devonian rocks in detail. The present chart is therefore the first comprehensive one to be offered.

ACKNOWLEDGMENTS

This chart has been compiled by the writer with the help of several colleagues and the close guidance of Dr. Edwin Kirk. Cooper is mainly responsible for the arrangement of the Middle Devonian and the Devonian of parts of western United States and of Gaspe, Quebec Swartz arranged the Lower Devonian columns of the Appalachians, Butts advised on the sections in the Southern Appalachians; Willard furnished the columns above the Onondaga for Pennsylvania; Caster examined and approved the columns for northwestern Pennsylvania and southwestern New York; Merriam aided in the preparation of the sections for the Eureka district, Nevada and California; Kindle furnished the columns for the Northwest Territories, Canada; Warren submitted sections and correlations for the Canadian Rockies; Warthin contributed notes on the Skunnemunk outlier; Mr. Wilson M. Laird kindly furnished new information on southwestern Pennsylvania; Dr. M. A. Stainbrook helped with a discussion of the Independence shale; Dr. G M Ehlers gave information on the black shale sequence of Michigan; Dr. W. L. Bryant helped with a discussion of the age of the Escuminac formation of Quebec. Dr. Winifred Goldring furnished a discussion of the Schoharie formation in New York. New information on the Devonian of Michigan and the Midwest is furnished by Cooper and Warthin. In the discussion accompanying the chart each contribution not furnished by Cooper bears the name of the contributor.

GEOGRAPHIC ARRANGEMENT

To show the facies of the sediments best the columns start with Alabama and Georgia and proceed northeastward along the Appalachians into the Catskill "delta" region southwest of Albany, New York, and then swing sharply westward to Lake Erie, southwest to northern Ohio, and south along the east flank of the Cincinnati Arch to southern Kentucky. From Ohio the sequence is extended to Michigan and the Midwest. This arrangement shows well the great extent of the red-bed facies and the passage westward of the normal marine sequence to black shales in the Midwest

The Midwest columns are made to adjoin those of Manitoba where the *Stringocephalus* zone is well developed. With this zone as the main tie the sections of the Great Basin, Canadian Rockies, and Northwest Territories are placed next to the Manitoba sequence. The columns for New England, Maritime Canada, and Gaspe Peninsula show a great development of the Lower Devonian and do not fit well at either end of the chart.

These might have been better placed adjacent to the New York columns, but this would have obscured the facies relations of the geosyncline.

DIFFICULTIES IN CORRELATION

Many of the correlations indicated are only tentative; the reader is warned not to accept the chart as a final or fixed expression of correlation; *it is a statement of existing knowledge only.* The major factors preventing accurate correlation at the present time are: (1) lack of detailed field work in certain areas, and (2) inadequate paleontology.

(1) Many Devonian areas of the United States and Canada, such as the Great Basin and other western areas, are known only superficially. Merriam's recent work has promoted better understanding of the Devonian of the West, but much still remains to be done. More comprehensive studies are needed on the Jefferson limestone of Montana and Wyoming, the Devonian of California, of New Mexico, and of southern Nevada. Stoyanow has made a significant contribution to the knowledge of the Devonian of Arizona.

Most of the Devonian areas in the Midwest are so well known that the problem there is for the paleontologist rather than for the stratigrapher and areal geologist. In the East a great blank exists in the knowledge of the Devonian of the Appalachians south of Maryland. The generalities of the outcrop belts are known, thanks to the mapping of Butts, but details of the stratigraphy such as facies relationships have never been worked out.

(2) If detailed correlations in this country are ever to be perfected many more fossils must be described. For years we have talked of Traverse faunas, Cedar Valley faunas, and others, but most of their species are yet unnamed. The writer has about 250 species of brachiopods alone from the Traverse group of Michigan, and most are new. Throughout the Appalachians and New York many important species are still undescribed. Settling of the Devonian-Mississippian boundary in the Appalachians and Allegheny Plateau awaits Caster's promised description of the Conewango and related faunas. The major outlines of the Devonian stratigraphy of Pennsylvania have been laid down by Swartz, Cleaves, and Willard, but the paleontology and unraveling of important faunal zones are still to be done. Only through such detailed work will exact correlation of formations of New York and Pennsylvania be possible. In the Midwest, redescription of the Cedar Valley (now being undertaken by M. A. Stainbrook), Callaway, Independence, Grand Tower, and St. Laurent faunas is urgent.

Including Walcott's paleontology of the Eureka district, Nevada, and the Territorial Survey reports, only a handful of fossils has been described from the West, and many are now known by eastern names. Most of the descriptions of species need revision.

STANDARD SECTION

The standard column of the Devonian of the United States is in New York where an almost unbroken and undisturbed sequence extends from the Coeymans limestone of the Helderberg stage into the Mississippian system. The only significant gap is that of the *Stringocephalus* zone which apparently pinches out to the southeast from Presque Isle County, Michigan. Nevertheless, the position of this zone is now known, and the upper Hamilton and Tully (Taghanic) are thought to represent its upper part. With this qualification the New York column is an excellent standard, and all the sequences of the Midwest and West have been compared to it.

DIVISIONS OF THE DEVONIAN

Ever since its establishment on this continent the Devonian system has been separated into a lower, middle, and upper part. This threefold arrangement unfortunately has become intrenched in the literature. As classified, the divisions are unequal, the Upper Devonian in general outweighing the other two, and the Lower Devonian being represented by the Helderberg and Deerpark stages only. The writer has long wondered whether these divisions of the American Devonian actually correspond with those of the European column. It seems wise therefore to divide the American Devonian into stages and reserve judgment on the parts to be placed in the lower, middle, and upper divisions.

According to present American classification the Helderberg and Oriskany constitute the Lower Devonian; the Onondaga and Hamilton form the Middle; and the remainder is included in the Upper Devonian. By comparison with the European column a somewhat different arrangement would include the Onondaga besides the Helderberg and Oriskany in the Lower Devonian. In Germany *Paraspirifer cultrijugatus* characterizes the Upper Coblenzian and Eifelian. The former, at the top of the Lower Devonian, seems to be the equivalent of the Onondaga, while the Marcellus correlates with the Eifelian rather than the Givetian where it has hitherto been placed. This is based on the range of *Paraspirifer* from the Onondaga through the Marcellus and the fact that the Marcellus assumes an Eifelian or "reef" facies in the Midwest and Canada. It is thus contended that the Onondaga with its "reefy" facies has been correlated incorrectly with the "reefy" Eifelian and that the shaly Marcellus actually correlates with the Eifelian. This is borne out by the fact that in Michigan the *Stringocephalus* zone rests not on the Onondaga but on the Dundee or calcareous equivalent of the Marcellus.

Essentially the same condition holds in Lone Mountain, Nevada, where a fauna suggestive of the Delaware limestone (*Martinia* zone of Merriam) underlies the *Stringocephalus* beds and overlies an Onondaga equivalent. If the Marcellus is Eifelian the Onondaga becomes Lower Devonian.

A major change is also necessary in the Middle Devonian: the transfer of the Tully to its old position at the top of the Middle Devonian. The Tully contains a Middle Devonian fauna and is actually the equivalent of the upper Givetian, not earliest Frasnian as hitherto claimed.

In subdividing the Devonian the writer is employing several new stage names. These stages represent groupings of formations chiefly on a faunal and paleogeographic basis. Seven of the names are new, but two are familiar group terms elevated to the rank of stage.

Helderberg stage.—The group term of this name is elevated to stage rank. The type section is in the Helderberg Mountains of New York.

Deerpark stage.—This stage includes the Oriskany and its correlates. The type section is in southeastern Deerpark Township on the west slope of the Shawangunk Mountains facing Port Jervis, Sullivan County, New York.

Onesquethaw stage.—This stage includes the sediments of the Esopus, Schoharie, and Onondaga and their correlates elsewhere on the continent. The type section is in the Helderberg Mountains west of Clarksville and facing the valley of Onesquethaw Creek in Albany County, New York.

Cazenovia stage.—This is a revival of the Cazenovia group of Conrad (1841) and Vanuxem (1842) as a stage term. The name as here used includes the strata from the top of the Onondaga to the base of the Centerfield, or the Marcellus and Skaneateles formations. The type section is in the township of Cazenovia, New York, where the entire sequence occurs.

Tioughnioga stage.—This includes the Ludlowville and Moscow formations which are closely related faunally and are mainly restricted to the eastern part of the Devonian geosyncline. Except for the Centerfield and its equivalents this stage is not represented in the Ohio and Mississippi valleys. The name is taken from the headwaters of Tioughnioga River in the south half of the Cazenovia quadrangle, New York. The interval embraced by the new name is from the base of the Centerfield to the base of the Tully limestone.

Taghanic (or Taughannock) stage.—This includes the sediments of the Tully and Geneseo with their correlates elsewhere on the continent. The type section is in Taghanic (Taughannock) Falls Park northwest of Ithaca, west side of Cayuga Lake, New York.

Finger Lakes stage.—This new name includes the strata from the top of the Geneseo shale to the base of the Chemung stage (Cayuta shale and sandstone and Grimes sandstone). The type section is in the Finger Lakes country of New York where these rocks are well exposed on all the larger lakes. The sequence thus includes the Genesee and Naples groups as conceived by Chadwick which are closely linked faunally.

Chemung stage.—The group term Chemung is here elevated to stage rank because of the widespread and distinctive character of the fauna. The stage has the same limits that Chadwick gave to the group.

Cassadaga stage.—This is a new name for the Canadaway and Conneaut groups which are faunally related to the Chemung and to each other. The name embraces the sequence from the Dunkirk shale to the Wolf Creek-Panama conglomerate. The type section is the headwaters and valley of Cassadaga Creek in the Dunkirk, Chautauqua, and Jamestown quadrangles, New York.

Conewango stage.—Above the Cassadaga stage (top of Conneaut) the fauna is distinctly different from that below, and the Conewango group term is therefore here elevated to that of a stage.

FACIES

GENERAL STATEMENT

The Devonian of New York State presents an excellent example of shifting facies, particularly in the post-Onondaga parts where the changes of sediment have been traced from red beds in eastern New York to black shales and limestone in Ohio. In general the change is from red sands and conglomerates to gray, fine-grained sandstones to dark siltstones, then to dark-gray and black shales, and finally to calcareous shales and limestones containing coral plantations and bioherms. The pattern of these shifts is now so well known that relationships not yet recognized in parts of the Appalachian geosyncline can be anticipated.

Following the Onondaga the great clastic flood was initiated by the Marcellus. In eastern New York the lowest Marcellus is a fine sandy gray shale which gradually passes into black shale and black limestone near Cayuga Lake. While the sands of the Mount Marion formation were being deposited in eastern New York, black muds of the Chittenango were forming in east-central New York. In late Marcellus time the first nonmarine beds, the Ashokan sandstone, appeared in eastern New York and were followed by red beds of the Skaneateles. The black shales continued to form throughout the rest of Hamilton time but appeared farther and farther westward and also higher and higher in the section so that black and gray mud facies occur in all the Hamilton stages. Obliquely above the black muds occur the silts and fine sands of the next facies. These siltstones are overlain obliquely by coarser sands that occupy the eastern part of the State and extend as far west as central New York. The red bed or nonmarine facies obliquely overlies the sandy facies and in the Hamilton extends westward only as far as Schoharie Valley.

A slight break in sedimentation occurs at the top of the Hamilton resulting from formation of a shoal on the site of Lake Erie probably in connection with movements of the Cincinnati arch. At any rate the movement is shown in a slight break traceable nearly to Schoharie Valley where all evidence of disconformity is lost. Following this break the Tully sediments show nearly the same facies shifts as the Hamilton. The calcareous facies is located between Canandaigua Lake and Chenango Valley. In Unadilla Valley, 10 miles east of Chenango Valley, the Tully is represented by clastics which in Butternut and Otego valleys are more than five times as thick as the limestones. In Schoharie Valley red beds of Tully age appear in the sequence, and conglomerates and red beds at the top of the Catskill Mountains facing the Hudson Valley are thought to represent this formation.

In post-Tully time the same sequence of facies occurred with the red beds pushing progressively westward until they reached the meridian of Canandaigua Lake. The black shale occupies most of the section between the Genundewa and the Gowanda. Fine-grained gray sands and siltstones lie between the reds and blacks.

West of New York, facies shifts are still to be recognized. The Helderberg stage and most of the Deerpark are overlapped by the Onondaga limestone, but in the Hamilton the black shales of New York interfinger with bluish Arkona clay shales and the Delaware limestone. In Alpena and Presque Isle counties, Michigan, the Marcellus of New York is represented by buff Delaware-Dundee limestones with "reefs" of the same age located in southwestern Ontario and the James Bay region. The Skaneateles dark muds of the Lake Erie shore are represented in Michigan by thick limestones, containing layers of calcareous shale (Bell to Alpena) and including large bioherms and biostromes of corals and stromatopores. Thus the limestones of Michigan represent the "reef" or shelf facies of the lower Hamilton. Southwest of New York the Logansport and Beechwood limestones of Indiana also represent the same shelf facies.

The Upper Devonian of New York contains no calcareous facies, and this phase of sedimentation is not encountered east of the Mississippi. West of the Mississippi calcareous shale facies and some Upper Devonian limestone occur in Iowa and Missouri, but the great development of calcareous Upper Devonian is in the Great Basin. It is therefore difficult to correlate the sands and shales of the Upper Devonian of New York with the limestones of the West.

These facies changes are also well developed in Pennsylvania where they have been described by Chadwick, Willard, Caster, and Laird. The same facies should also be well represented in the Appalachians southwest of Pennsylvania but have never been worked in detail. The outcrop of the Devonian cuts east across New York into the red beds southwest of Albany. It then turns southwestward and runs along the Catskill Front. Between Albany and Napanoch, New York, the two upper stages of the Hamilton are represented by nonmarine beds, but, to the southwest, marine beds dominate the section which is wholly marine from Port Jervis southwestward into Pennsylvania, Maryland, Virginia, and West Virginia. Red beds occur in the lower part of the Upper Devonian of the Delaware River sequence northwest of Port Jervis but, continuing southwest or toward the seaward part of the geosyncline, marine beds are more frequent and finally occupy the entire sequence. The Finger Lakes stage is mainly marine across Pennsylvania, Maryland, Virginia, and West Virginia. The reds of the higher Upper Devonian give way to marine sediments in western Pennsylvania, and representatives of the Canadaway and Conneaut groups are known in western Virginia. It is thus clear in the Appalachians that, as the folded belts bring up sediments representing more seaward parts of the geosyncline, the sequence becomes more and more like that of the New York outcrop. In other words the Appalachian sequence when once unfolded should in its major features and its faunal aspect duplicate the New York succession.

BLACK SHALE

Much discussion has centered around the age of the black shales of the Midwest and South; one school of thought claims that these shales are Devonian, while the other believes they are Mississippian. Because of the established facies relationships of the Devonian of New York and Pennsylvania, the writer believes that much of the black shale of Ohio, Indiana, Kentucky, Illinois, and Michigan is Devonian. It is still a problem where to assign some of the black shale of the South.

Portions of the black shale containing *Lingulipora, Schizobolus, Styliolina, Hypothyridina,* and *Leiorhynchus quadricostatus* are generally conceded to be of Devonian age. Such deposits occur in Virginia, Indiana, Kentucky, Tennessee, Oklahoma, Arkansas, and possibly Alabama. In New York, *Schizobolus* occurs in the upper beds of the Hamilton and becomes extinct in the Naples group. Besides occurring in the Hamilton and Geneseo, it is reported from the Marcellus in Virginia. This particular black shale (Millboro) in Virginia may conceivably represent not

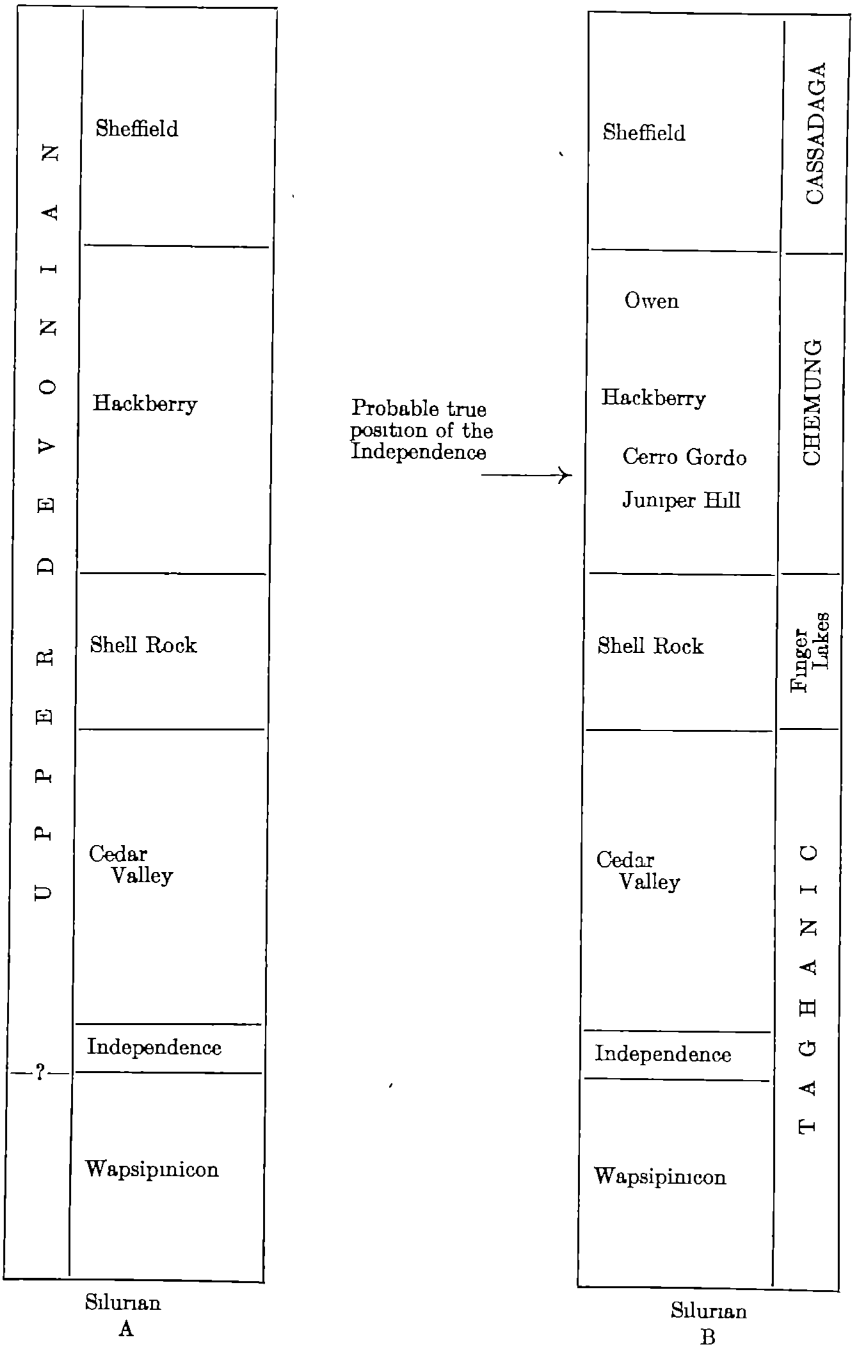

FIGURE 1.—*Diagram showing contrasting views of the interpretation of the Independence shale*

A. Stainbrook; B Cooper and Warthin

only the Marcellus but the entire Hamilton. In the Midwest and in Tennessee, *Schizobolus* has generally been assigned to the Geneseo, but the presence of *Hypothyridina* in similar shales suggests that perhaps they should be assigned to the Tully. As both the Tully and Geneseo are here placed in the Taghanic stage this question does not greatly affect accurate correlation.

The real problem of the black shale is not with the beds carrying *Schizobolus,* but with the overlying strata. In general this portion has few fossils, and, except for the Kinderhookian (Hamburg) fauna at the top, has yielded no diagnostic specimens. The brachiopod *Barroisella subspatulata* has often been regarded as a Mississippian guide fossil, but in the New Albany shale it occurs in or near a green band not far above the Geneseo equivalent which constitutes the lower 10 feet of shale In Michigan the writer collected this species in a patch of green shale on the floor of the shale pit at Paxton, 8 miles west of Alpena, below concretions containing the Devonian goniatite *Tornoceras.* As this goniatite is unknown in the Mississippian, the Devonian age of *Barroisella subspatulata* is fixed. This common brachiopod has been reported also from the Mountain Glen shale of Illinois.

When the facies relationships of the Upper Devonian are considered it would be expected that the great mass of sediments of the Appalachian geosyncline would be represented by black shales in the off-shore and pre-shelf region. Study of the Hamilton and Upper Devonian corroborates this. In northern Ohio the Chagrin shale represents the silty facies of the Conewango group and part of the Conneaut, and, west of the Chagrin area, black shale appears until finally the Huron shale is thought to represent the black shale facies of the Chagrin. The Ohio shale is believed to equal the Huron and "Portage" black shale.

Passage of the Upper Devonian sediments to black shale is accompanied by thinning of the sequence. For example, the Oatka Creek shale of the Marcellus stage is 50 feet thick in Erie County, New York, while its coarser equivalent in the Catskill region is about 1300 feet thick. The Upper Devonian black shales thin to the west, except locally, but it cannot yet be proved whether the outlying strata equal all the Upper Devonian or whether there is overlap. In the latter case the beds farthest out would be youngest. Local thinning by unconformity at the top is known, but it has not been demonstrated as universal over the range of the black shale.

In southwestern Virginia the gray Brallier shale represents part of the Naples group and possibly some of the Chemung as restricted, but black fingers have been detected in this formation which Butts (oral communication) thinks may represent the incoming of the "Chattanooga" facies. This is possible, but the southern or type Chattanooga has not been dated so low in the Upper Devonian.

Although direct fossil evidence of the age of the black shales has not yet been found the presence in a number of areas of a basal Kinderhook fauna gives a definite upper limit to the Devonian shale. The fauna in question is that of the Hamburg oölite that occurs in the Glen Park limestone in Calhoun County, Illinois, and can be identified at a number of localities east and west of Calhoun County. To the west the Glen Park limestone contains this fauna. To the east it has been seen at the top of the New Albany shale (Huddle, 1933) and it occurs also in the Bedford shale of Ohio. Foerste (1909) traced the Bedford fauna into eastern Kentucky. Savage and Sutton (1931) report it in south-central Kentucky, and J. H. Swartz (1924) describes it at Eulie, Sumner County, central Tennessee. The fauna reported by J. H. Swartz (1929) from the Olinger shale of eastern Tennessee may be a far-removed outpost of the same assemblage.

The Devonian or Mississippian age of the Hamburg fauna has not yet been settled. Although originally described as Mississippian (Weller, 1906) a Devonian age was claimed by Girty (1912) for its equivalent in the Bedford shale. More recently Branson and Mehl (1938) and Branson (1938) champion the Devonian claim. The majority of species do have unmistakable Devonian affinities and include some familiar types: *Nucleospira* and *Atrypa* of the brachiopods and *Modiomorpha, Sphenotus,* and *Cypricardella* of the pelecypods. In a long list of Devonian forms the few Mississippian elements are *Syringothyris* (not always Mississippian), *Spirifer marionensis,* and *goniatites.* Total absence of true productids is significant. At present the fossils have not been sufficiently well studied or illustrated to date the beds definitely. Should the fauna prove to be Devonian the black shales will be Devonian; but if its age remains Mississippian, as it is by definition, the lower Kinderhook being part of the type Mississippian, the fauna will serve as an excellent ceiling for the Upper Devonian.

In the South the age of the Chattanooga has been scrutinized by J. H. Swartz (1924) who describes a section near Apison, 16 miles east of Chattanooga. Above 10 feet of black shale in this section about 2 feet of dark-gray shale contained *Lingula irvinensis, Rhipidomella, Chonetes acutiliratus, Schuchertella,* and *rhynchonellids* which identify it with the Bedford-Berea wedge. Above this layer *Lingula melie* indicates the Sunbury shale. According to Swartz the shale below the Bedford-Berea is of Cleveland age, here assigned to the Devonian In the Big Stone Gap region Swartz (1929) shows the Big Stone Gap shale of Stose to be divisible into the Cumberland Gap shale at the base, the Olinger shale, and the Big Stone Gap shale, restricted, at the top. The Olinger contains a fauna suggestive of the Hamburg. These shales have been correlated with the Chattanooga, and Swartz concludes that the Chattanooga shale in Tennessee and Virginia is Mississippian "with the possible exception of the lower part of the Cumberland Gap member." Inasmuch as this member interfingers with the Olinger it is contemporaneous with it.

In Oklahoma, Missouri, Mississippi, and Arkansas thin black shale formations occur that have been loosely called Chattanooga. Those containing the *Schizobolus* fauna are definitely Devonian and probably are more accurately called Trousdale, the Geneseo black shale of Tennessee. The other black shales called Chattanooga may be the feather edge of the Upper Devonian, but the point cannot be proved. In Alabama the writer has identified *Lingulipora,* a Geneseo brachiopod, in greenish rocks associated with the black shale and previously called Chattanooga.

In summary it may be said that according to the writer's view most of the black shale discussed in the Midwest and Appalachian region represents a black shale facies of the Upper Devonian underlying a fauna of highest Upper Devonian or basal Kinderhookian age. These black shales may represent the feather edge of the thinning Upper Devonian, thickening locally on the flank of the Cincinnati arch as shown by Caster (1934) in the case of the Cleveland shale. Furthermore the Bedford-Hamburg shows a facies change in harmony with that of the Devonian—*i. e.* shale in Ohio and Indiana and limestone in Illinois and Missouri.

IMPORTANT NEW CHANGES IN CORRELATION

For emphasis the important changes in Devonian correlation and thought proposed in the chart and accompanying text are as follows:

(1) Grouping of the Devonian into 10 stages.

(2) Assignment of Heppel, Causapscal, and Gaspe formations to the lower Onesquethaw stage rather than to the Hamilton and Naples groups.

(3) Assignment of the Tully and Gèneseo formations to the Middle Devonian, Taghanic stage.

(4) Correlation of the upper Traverse of Michigan with the Tully and Cedar Valley formations.

(5) Fixing of all formations correlating with parts of the Hamilton group in the Mississippi Valley as Centerfield or older.

(6) Correlation of the Logansport limestone of Indiana with the Hamilton Beechwood and Four Mile Dam limestones rather than with the Jeffersonville.

(7) Correlation of the Lake Church of Wisconsin and the Abitibi River limestone of James Bay with the Delaware-Rogers City limestone of Presque Isle County, Michigan.

(8) Assignment of the *Paraspirifer* zone of the Silica shale to the Marcellus and the upper Silica to the Ferron Point-Genshaw formations.

(9) Determination of the Mount Marion of the Catskill region, New York, as a partial equivalent of the Chittenango; placing the Ashokan in the Marcellus; correlation of the Bakoven with the Union Springs and the Stony Hollow member with the Cherry Valley limestone.

(10) Placing of Percha shale high in Conewango stage.

(11) Assignment of the Jefferson formation and equivalents to Finger Lakes and Chemung stages.

(12) Recognition of Shell Rock elements in lower Devils Gate formation.

DISCUSSION OF FORMATIONS

NAMES NOT APPEARING ON CHART

Not all valid names appear on the chart; many are too local to be fitted in. Others are so isolated from main sections as to require too many additional columns thus making the chart too unwieldy. For completeness the more important of these names are listed below. These are arranged by the States in which they occur, and their major features can be determined by reference to the *Lexicon of geologic names of the United States* (Witmarth, 1938). It is not claimed that these names plus those appearing on the chart include all Devonian formation names. Synonyms are not included, but some of the names in the list and on the chart may ultimately prove to be synonyms. Named igneous and subsurface formations assigned to the Devonian in general do not appear on the chart or in the list.

During the preparation of the chart it became apparent that many formations were incorrectly correlated. The writer believes that in reassigning these formations it is essential to give the reasons for the shifts. Too many charts with new and far-reaching correlations unsupported by proof have been published and have resulted in no end of confusion. Consequently all new changes on the chart are explained. In some instances the reasons given may not satisfy the reader or other Devonian students; nevertheless the evidence, good or bad, is given. To keep the length of these notes within reasonable bounds information relating to lithology, thickness, and the like has not been repeated; this information can be obtained from the *Lexicon.* Very little information pertaining to paleontologic evidence for correlations is given in the *Lexicon;* therefore the writer's remarks contain mostly paleontological information.

- Alaska
  - Takotna
  - Vallenar
  - Tonzona
- Arizona
  - Morenci
  - Sycamore Creek
  - Temple Butte—Chemung or Canadaway
- Canada
  - Camp Creek
  - Famine (see text)
  - Horn River
  - Knoydart
  - Mount Charles
  - Nictaux
- Illinois
  - Hoing (subsurface)
- Indiana
  - New Chapel chert (in Silver Creek formation)
- Iowa
  - Montpelier
  - Raymond Quarry
- Kansas
  - Ediger (subsurface)
- Maine
  - Mapleton
- Maryland
  - Avilton
- Massachusetts
  - Vernon
- Missouri
  - Auxvasse Creek
  - Noel
- Montana (all Upper Devonian)
  - Ermont
  - Monarch
  - Union
- Nevada
  - Combs
  - Lamoureux
- New York (all Upper Devonian)
  - Barryville
  - Keuka
  - Long Beards Riffs
  - Mount Herman
  - Pope Hollow
  - Tunangwant
  - Wittenberg
- Ohio
  - Bellepoint
  - Delhi
  - Dublin
  - Eversole
  - Hillsboro (see Detroit River)
  - Klondike
  - Marblehead
  - Venice
  - West Jefferson
- Oregon
  - May Creek
- Pennsylvania
  - Armenia
  - Bimber Run
  - Cascade
  - Coudersport
  - Dutchmans
  - Franklindale
  - Hosmer Run
  - Kingsley
  - Lanesboro
  - Luthers Mills
  - Millers
  - New Milford
  - North Warren
  - Piney Ridge
  - Roystone
  - Saxton
- Tennessee
  - Hilton
  - Swan Creek
- Washington
  - Orcas (see text)
- West Virginia
  - Rowlesburg
  - Saxton

EXPLANATION OF CHART

In using the chart the reader must bear in mind the limitations of portraying correlation by this means. In the first place, many of the columns represent sections along meridians rather than at the place named. For example, the Canandaigua Lake column includes more names than those representing rocks appearing on the shores of the lake, the additional ones occurring some distance to the south. Secondly, relative thicknesses are not represented by the size of the compartments separating stages and formations. The space allotted to the stages was arbitrarily planned, and the space assigned to formations in general depends on the number of names appearing in the compartment. The column for Matapedia, Quebec, illustrates the point. The Causapscal, Heppel, and Four Mile Brook formations aggregate several thousand feet, yet they are believed to belong in the Onesquethaw stage of New York which includes a thickness of less than 500 feet of rock.

In a few instances columns are composites of the Devonian formations appearing in a State or region. The formations are arranged according to their position relative to the standard column although they may not appear in continuous sequence in the field. This is true of the column representing the Brisco Range and Windermere District.

In many instances correlations indicated are only suggestive. The Jefferson limestone of northwestern Montana now known to be Upper Devonian is divided into five members as shown by its fossiliferous middle member and is correlated with the New York Chemung. The members above and below the Coopers Lake limestone are also assigned to the Upper Devonian but without conclusive paleontologic evidence. This sort of difficulty also arises in the correlation of units in two separate regions, the limiting members being definitely correlated but the inner members of uncertain affinities. In such an instance the inner members are accommodated to the available space, and definite correlation not indicated. An example is the Ithaca formation of New York consisting of six members of uncertain relationships.

At the extreme left of the chart the ranges of some important fossils are given to help the reader understand the correlations. In general the attempt was made to record the ranges of genera mentioned in the text or those sufficiently restricted and easily recognized to make short segments of the column readily identified.

Abitibi River formation.—Redefined by Kindle (1924) to eliminate the beds containing *Hypothyridina* (now called Williams Island limestone) about 8 feet thick. As thus redefined this limestone is regarded as of Delaware and Rogers City age because recorded lists of fossils show no species entirely restricted to the Onondaga where the formation had been hitherto placed. The species *Gypidula comis, Martinia subumbona,* and *Delthyris consobrina* (possibly *Spirifer lucasensis*) listed by M. Y. Williams (1920) are unusual in or absent from the Onondaga but are common Delaware or Rogers City fossils. *Schizophoria* and *Athyris,* common in the Abitibi River limestone, are present in Onondaga rocks but more abundant in the Delaware. None of the species listed by Savage and Van Tuyl (1919) from the Abitibi River lends any support to their suggested correlation with the late Devonian of Iowa.

Alpena limestone (A. S. Warthin and G. A. Cooper).—Further restriction of this name is necessary after more detailed studies by the authors. The lower 15 feet contains a Genshaw fauna and has been placed with that formation. The next division, called Newton Creek limestone, consists of 25 feet of yellowish to brown crystalline limestone containing oil in cavities and fossils. Large numbers of *Pentamerella, Camerophoria,* *Charionella, Cranaena,* and color-banded cephalopods are present. The succeeding limestone, to which the name Alpena is restricted, consists of 79 feet of light-gray to white limestone. In the upper 40 feet large bioherms of *Prismatophyllum* and *Stromatopora* are conspicuous. Above the Alpena (restricted) occur 8 feet of blue clay shale, the Dock Street Clay of Grabau (1902), a lens confined to the east side of Alpena, and classified with the Alpena limestone. Overlying the Dock Street and the Alpena and faunally related to them is the Four Mile Dam limestone, 8 feet thick at the south end of the quarry of the Thunder Bay Quarries Company. The type section is at Four Mile Dam where the formation is represented by a bioherm containing many elements of the Centerfield fauna of New York.

Alto formation.—Correlated with the Cornell and Sherburne formations of New York because of the reported presence of *"Reticularia" laevis* (Hall) (*see* Savage, 1920.), but may actually belong in the Taghanic stage because of indicated Tully affinities (Savage *in* Jillson, 1931).

Amherstburg dolomite.—One member of the Detroit River series containing among other Onondaga types *"Spirifer" divaricatus* Hall pointing to a position in the lower part of the Onondaga (Schoharie?). (*See* Detroit River.)

Amity shale.—Low in the Conewango group of Chadwick. Contains the peculiar spiriferoid *"Cyrtia" alta* Hall which also occurs in the lower part of the Chagrin shale of Ohio. *"Cyrtia" alta* does not belong to *Cyrtia,* a Silurian genus, because it differs in the structure of the beak, apical plate, form of the valves, and ornamentation. The Silurian *Cyrtia* possesses a perforate arched deltidium, whereas that of *"Cyrtia" alta* is an imperforate plate not arched above the interarea and is thus like that of *Syringothyris. "Cyrtia" alta* belongs to Kindle's genus *Syringospira* (1909, p. 28) from the Percha shale of New Mexico which possesses the extremely elongated ventral palintrope, apical plate and costellate fold and sulcus exactly like the New York and Ohio species. (*See* Percha shale, Ouray limestone.)

Antrim black shale.—In Alpena County, Michigan, the lower part of this shale above the Squaw Bay formation contains *Styliolina* in abundance, and this fossil is also known in the black shale just west of Beebe School, southwest of Afton, Cheboygan County. It has been generally believed for some years that the black shale above the *Styliolina* beds is Mississippian. Against this view is the fact that Devonian fossils have been collected in the great shale pit at Paxton, 8 miles west of Alpena. The writer discovered *Barroisella subspatulata* (Meek and

Worthen) in patches of green shale on the floor of the quarry, and Ehlers (personal communication) and his students collected the goniatite *Tornoceras* in a concretion higher in the quarry. This is a typical Devonian fossil as yet unknown from the Mississippian.

In the supposed Waverleyan beds exposed on the shore of Lake Michigan, a mile north of Norwood, Cooper and Cloud with Ehlers and party found *Styliolina, Paracardium,* and *Buchioloa* in the hard greenish layers at the base of the black shale. These fossils indicate a horizon probably higher than the Squaw Bay limestone, but nevertheless low in the Upper Devonian. The black shale above the *Paracardium* beds contains Devonian conodonts (Ehlers, personal communication).

Arkona shale.—Represents final passage of the dark Levanna shale of New York to a light-colored calcareous shale. Black fingers of Levanna type, containing *Leiorhynchus,* occur near the top of the Arkona. Fossils are most like those of the Chenango sandstone of the New York Hamilton, but some Michigan elements suggestive of the Bell and Ferron Point shales are present.

Ashland limestone.—Rensselandia [*Newberria*] zone of central and northeastern Missouri.

Ashokan sandstone.—Recognized in the Catskill region northeastward from Kingston nearly to Albany and southwestward to the central part of the Rosendale Quadrangle. The Ashokan is for the most part composed of poorly fossiliferous, nonmarine sandstone containing a few plants and fishes. In the Coxsackie Quadrangle marine fossils appear sparingly toward the seaward side of the mass. It overlies the Mount Marion formation of the Marcellus stage which contains abundant *Paraspirifer.* An upper Marcellus age is suggested because the thickness of the Ashokan combined with that of the Mount Marion and Bakoven is about 1300 feet, a thickness about equal that of the Marcellus of the Schoharie Valley (1000 feet) plus the normal increase of thickness expectable between Schoharie Valley and Catskill.

Bakoven shale.—Represents the easternmost known facies of the Union Springs shale. In Cayuga Lake region the Union Springs is a black limestone, but in eastern central New York it is a sooty black shale. In the Catskill region the Union Springs has become a dark sandy shale abounding in *Paracardium* and *Buchiola.*

Bear River formation.—As some uncertainty exists regarding this formation the original description is reproduced:

> ". . . The only marine Devonian known to date in Nova Scotia is found in Annapolis county in the vicinity of Bear River and Nictaux where it consists for

> the most part of dark gray and green and brown shales or reddish sandstone or arenaceous limestone, considerably squeezed and altered, constituting the *Bear River* formation *Pleurodictyum problematicum* is one of the characteristic species of this horizon which is evidently lower or Eo-Devonian . . ." (Ami, 1900, p 206-207).

Beartooth Butte formation.—Placed in the Oriskany-lower Onondaga (=Schoharie) position opposite that portion of the Nevada limestone, the only other known lower Devonian of the West.

Beauvais sandstone.—Fossils such as *Leptaena rhomboidalis, Schizophoria striatula, Spirifer varicosus, Paracyclas elliptica,* and *Proetus* cf. *P. haldemani* reported by Croneis and Hoffman (1931) suggest relationship to the Silver Creek limestone of Indiana. In addition to these fossils poor impressions of the brachiopod *Rensselandia claypolei* (so-called) have been taken. The presence of this genus indicates a Middle Devonian age. The Beauvais lies between a possible Delaware equivalent, the upper Grand Tower, and a probable Skaneateles equivalent, the St. Laurent limestone. The formation has therefore been placed as a facies of the St. Laurent and a possible correlate of the Rogers City limestone which contains *Rensselandia* and occurs in the same position between the Delaware and the Skaneateles.

J. M. Weller (1939) suggested correlation of the Beauvais with the Dutch Creek sandstone because they are lithologically alike, but the two are totally unlike in fauna, and the writer therefore regards the correlation as wholly untenable.

Bedford shale.—Contains a fauna which, according to Caster's lists (1934, p. 158-164), the writer correlates with the fauna of the Hamburg oölite of Illinois, the Glen Park limestone of Missouri, and the marine fauna at the top of the New Albany (Huddle, 1933). The age of the Bedford and its correlates has not yet been settled, a number of geologists regarding it as Devonian, while others hold it to be Mississippian. The date of the formation cannot be settled here; nevertheless it can be stated that this fauna is widely distributed and occurs very close to the Devonian-Mississippian boundary and is the suggested upper limit for the Devonian on the chart.

Beechwood limestone.—Contains the chief elements of the Centerfield fauna of New York: *Pentagonia biplicata, Camerophoria, Camarospira, Cyclorhina, Centronella, Vitulina,* and *Eleutherocrinus cassedayi* (Cooper and Warthin, 1942).

Bell shale (Warthin and Cooper).—Hitherto classified with the Marcellus, this shale lies on the deeply eroded surface of the Rogers City limestone which contains the *Stringocephalus* fauna. Because the Rogers

City overlies the Delaware formation of the Marcellus and underlies a Centerfield equivalent, the Bell is proved to be of early Skaneateles age.

Bellvale sandstone.—According to Willard (1937, p. 271-272) this sandstone contains a Hamilton fauna that correlates with the Marcellus because the formation overlies the Cornwall shale of Onondaga age. Presence of Portage equivalents is most unlikely. (*See* Cornwall shale.)

Bernardston formation.—Most students of this formation agree to a Devonian age of at least part of it. Whitfield (1883, p. 368) suggested a Silurian age for the lower limestone and a probable Middle Devonian age for the overlying sandstone. Clarke (*in* Emerson, 1898, p. 259) on the other hand stated a late Devonian age for the shaly quartzites.

According to Emerson (1898, p. 264, Fig. 17) the fossiliferous rocks consist of limestone containing *Favosites,* cyathophylloids, and large and small crinoid stems. Over this lies a bed of magnetite up to 3½ feet thick on which occurs a calcareous sandstone containing much distorted impressions of fossils. The writer saw no identifiable specimens from this sandstone.

Clarke's contention that these beds are late Devonian is untenable because: (1) According to present views, all late Devonian rocks are absent or of an entirely nonmarine facies on the Catskill Front, precluding the presence of marine Upper Devonian on the east side of the Hudson Valley. The main mass of the Catskill Front is now known to be of Hamilton and Taghanic age. The known marine Devonian that extends across the Hudson Valley ranges from Helderberg (Coeymans) to Onondaga. It is among these formations that one would expect to find relations with the Bernardston beds. (2) New fossil evidence in the form of a fairly well preserved *"Spirifer"* from the magnetite bed gives evidence for an Onondaga age. This *"Spirifer"* is of an unusual type now known only in Onondaga and Hamilton rocks, and the specimen itself is not wholly typical of the group to which it is here assigned. It is a completely costate *Spirifer* such as *S. venustus* and *S. divaricatus.* Because the specimen is an impression of the interior none of the finer ornamentation is preserved. This ornamentation is important in the definition of the genus and consists of zigzag concentric lamellae bearing a row of fine spines on the edges. The Bernardston specimen is most suggestive of *S. divaricatus* or *S. grieri* because of its strong costae but differs in showing no divarication of the ribs at their anterior ends.

Boule limestone.—Raymond (1930, p. 294-296, 300) applied this name to his zones 3 and 4 of the Roche Miette section. Zone 3 contains *Ambothyris* ("non-plicate *Cyrtina*") and *Gruenwaldtia* ("small *Atrypa*") like those of the Independence shale of Iowa and the "Percha" shale of the Sacramento Mountains of New Mexico.

Brallier shale.—In southwestern Virginia fingers of black shale have been detected in the Brallier which, according to Butts (personal communication) represent the appearance of the Big Stone Gap shale (of Stose). The latter has been correlated with the Chattanooga at its type locality.

Caballos novaculite.—Aberdeen (1938) studied the radiolarian fauna of this formation and stated that two of the species showed a "slight similarity to radiolaria described from the Tamworth series (Devonian) of Australia." Correlated with Camden and lower Arkansas mainly on lithology.

Callaway limestone.—Correlated with the upper part of the Cedar Valley limestone—upper Rapid, *Atrypa waterlooensis* zone to Coralville —following Branson (1924, p. 2) and Stainbrook (1935a, p. 260). Correlation based on abundance of *Cranaena,* presence of *Camerophoria gregeri* (Branson), large *Schizophoria,* and *Syringothyris*-like spiriferoids. The sandstone at the top of the Cedar Valley in Calhoun County, Illinois, contains Callaway types.

Camden chert.—Until now the Camden fauna has been recognized outside of Tennessee only in the Clear Creek chert of Illinois, but published lists and collections studied by the writer from the Southern Appalachians and West Virginia indicate that the Camden fauna is more widely distributed than hitherto believed. The fauna of the Huntersville chert and most of the so-called Onondaga of the South is of Camden age. The Camden itself is correlated with the Schoharie as explained below.

Canadaway group (extracted from G. H. Chadwick letter to Cooper, Jan. 25, 1941).—Changes are imminent that will invalidate Canadaway and Conneaut or require their redefinition, with probably another group introduced between them. The former correlation was:

	LAKE ERIE	GENESEE-OLEAN
Conneaut	Chadakoin ("Chemung pink")	Haymaker (Chadakoin "pink")
		Hinsdale
	Volusia (Girard of NW Pa.)	Cadiz (Volusia)
		Cuba
Canadaway	Northeast	Machias
	Shumla	
	Westfield	Rushford
	Laona	
	Gowanda	Caneadea (Gowanda)
	Dunkirk	Dunkirk (Canaseraga)

The Gowanda undoubtedly includes more than the Caneadea, and there is reason to believe that the Cuba is the Laona, a correlation indicating identity of Hinsdale and Shumla. The Haymaker should then be equivalent to part, though probably not all, of the Northeast shale. In fact the Cattaraugus as well as the Haymaker may correlate with the Northeast shale.

The Canadaway paleontologically should terminate upward at the base of the Cuba sandstone. A similar sharp faunal break comes at the base of the Laona sandstone. If the Cuba and Laona are identical, as now seems likely, then the group from the Cuba up through at least the Haymaker is not Conneaut which is defined in eastern Ohio where it includes the Girard and Chadakoin only. Chadwick suggests limiting the Canadaway to the Dunkirk and Gowanda and introducing a new group name for the beds between the Canadaway and Conneaut, as follows:

	Lake Erie	Genesee-Olean
Conneaut group	Chadakoin Girard	
Unnamed group	Northeast Shumla Westfield Laona	Haymaker Hinsdale Cadiz Cuba
Canadaway group	Gowanda Dunkirk	Caneadea Dunkirk

Canutillo formation.—Named by L. A. Nelson (1937) for black shale, sandstone, and gray limestone in the Franklin Mountains north of El Paso, Texas. This formation is assigned to the middle Devonian (Nelson, 1940).

A collection of fossils from opposite Vinton, Texas, sent to Dr. G. H. Girty and now in the National Museum contains the brachiopods *Schizobolus, Meristella, Mucrospirifer,* and *Leiorhynchus.* The presence of *Schizobolus* suggests correlation with the Geneseo shale of New York, the Trousdale shale of Tennessee and Arkansas, and the lower 10 feet of the New Albany shale of Indiana. This correlation, however, must be regarded as tentative because the complete fauna of the Canutillo formation has not been published. Moreover the possibility of the Canutillo formation equaling part of the Sly Gap formation should not be overlooked. The Sly Gap contains *Leiorhynchus* according to Stainbrook (1935b) and Laudon and Bowsher (1941) in a black shale. It is thus possible that the Sly Gap includes the Canutillo.

Cap Barré limestone.—Dated as New Scotland by the presence of the strange trilobite *Dicranurus limenarcha* Clarke, although a possible Coeymans age should not be overlooked. Alcock (1935, p. 67) includes this limestone in the Murailles formation because it cannot be recognized west of the coast.

Cap Bon Ami limestone.—Contains a few Helderberg species. Overlies the St. Albans formation of New Scotland age and underlies the Grande Grêve limestone of Oriskany age. Therefore tentatively assigned to the Becraft and correlated with the Dalhousie formation.

Causapscal formation.—Crickmay (1932, p. 376) suggests a correlation with the Grande Grêve limestone of Gaspe. The presence of *Eodevonaria, Leptocoelia,* and *Spirifer* cf. *S. modestus* are suggestive of the Schoharie and the Camden chert.

Cedar Valley limestone (Cooper and Warthin).—The age of the Cedar Valley has long been a moot question. The presence of *Hypothyridina* in the lower Cedar Valley led Thomas (1924, p. 408) and Weller (1909, p. 265-266) to correlate it with the Tully of New York and place the formation in the Upper Devonian. Stainbrook (1935a, p. 255) on the other hand believes the Cedar Valley is late Devonian because a Chemung fauna, the Independence shale fauna, underlies it.

Hypothyridina has long been a misleading fossil (*see Hypothyridina*); its supposed basal Upper Devonian position has been accepted almost without question in this country, and the presence of the genus has been used to outweigh all other evidence. Actually it occurs well down in the Middle Devonian in Europe and Asia; consequently its mere presence does not signify an Upper Devonian age.

The known fauna of the Cedar Valley contains no fossils of unequivocal late Devonian age. Not only are late Devonian guides absent, but the majority of the genera are those most commonly found in Onondaga to Tully strata, such as *Stropheodonta, Pentamerella, Leptostrophia, Megistrocrinus, Stereocrinus, Pentremitidea, Nucleocrinus,* and *Prismatophyllum.* Although some of these genera are also known in the Upper Devonian, they are most abundant in Middle Devonian rocks, and often their species make up a large percentage of the fauna. The following Cedar Valley genera are unknown from late Devonian rocks: *Rensselandia, Charionella, Billingsastraea,* and *Nucleocrinus.* The predominant aspect of the Cedar Valley fauna is that of the Middle Devonian. Significant late Devonian elements absent from the Cedar Valley are: *Cyrtospirifer, Manticoceras, Ambothyris, Strophonelloides, Pugnoides.*

The evidence for the age of the Cedar Valley based on the position of the Independence shale also leads to difficulties. This shale contains an undoubted Chemung fauna, but Stainbrook (1935a, p. 252) believes it to be stratigraphically below the Cedar Valley. Savage (1920, p. 179-180) believes that the Independence represents fillings of caverns in the Cedar Valley limestone and is therefore younger than the formation it seemingly underlies. Outcrops of the Independence are few, small, and scattered, and all those seen by the writers afforded evidence so equivocal that its postulated position below the Cedar Valley is doubted.

To emphasize the Middle Devonian age of the Cedar Valley it is important to recall that Stainbrook (1935a, p. 260) has already correlated it with unquestionable Middle Devonian rocks. According to him "The Cedar Valley is faunally related to the upper part of the Traverse beds as developed at Partridge Point and probably is of the same age." The beds at Partridge Point (Thunder Bay formation) underlie the Squaw Bay limestone of earliest Genesee age (containing the goniatite *Koenenites*) and correlated with the Genundewa limestone (basal Upper Devonian) of New York.

Not only the age of the entire Cedar Valley is difficult to establish, but the dating of some of its parts presents problems. At present the Solon or lower member is linked with the main body of the formation by the common presence of the unusual crinoid *Stereocrinus*. Cooper and Cloud (1938) suggested the correlation of the Solon (then called Linwood) with the Rogers City limestone (*Stringocephalus* zone) because *Rensselandia* occurred in both. Stainbrook (personal communication) has convinced the writers that the Solon cannot be divorced from the Cedar Valley as a whole.

The Mineola limestone of Missouri and most of the Rapid member of the Cedar Valley fall opposite one another on the chart because of correlation of the overlying Coralville member with the Callaway of Missouri and of the underlying Solon with the *Rensselandia* zone of Missouri. The upper zone of the Rapid formation (*Atrypa waterlooensis* zone) probably should be correlated with part of the Callaway.

Centerfield formation (Cooper and Warthin, 1942).—Contains the most widely distributed known faunal assemblage of the Hamilton, extending across New York and part of Pennsylvania into Ontario, Michigan, Indiana, Ohio, Kentucky, and possibly Illinois and Arkansas. This easily recognized fauna has made subdivision of the Hamilton outside of New York possible and intelligible. The most important genera and species of the Centerfield are: *Favosites turbinatus, Prismatophyllum, Megistocrinus, Dolatocrinus, Eleutherocrinus, Camerophoria, "Spirifer" sculptilis, "S." venustus, Camarospira, Pentagonia biplicata, Cyclorhina nobilis, Parazyga hirsuta, Centronella impressa,* and *Vitulina pustulosa.*

Smith (1935, p. 43) has suggested that the Centerfield formation be placed in the Skaneateles stage because the two are transitional on the Skaneateles Quadrangle, a condition that persists eastward to the east side of Chenango Valley. In western New York from Canandaigua westward and in Ontario the Centerfield is set off from the underlying shales by a sharp lithologic change, clearly disconformable. It is not impossible for the Centerfield to be transitional to the Skaneateles in the east and disconformable in the western part of the State. Usually in the section from Skaneateles eastward the lower part of the Centerfield contains, or is underlain by, cross-bedded sands that may in themselves represent breaks, or the two may be falsely transitional because of reworking of older Skaneateles sediments. The faunal break with the Skaneateles is as striking in the eastern part of the State as it is in the western part. The writer therefore sees no reason to transfer the Centerfield from its traditional position at the base of the Ludlowville.

Chagrin shale.—Represents westward facies of the Conewango stage including most of the Chadakoin of New York and Pennsylvania. According to Caster (1934, p. 149) some of the divisions of the Conewango sandstones and shales of western Pennsylvania can be identified in northeastern Ohio. The faunas of the Panama-LeBoeuf conglomerate can be recognized by the presence of *Ptychopteria* and *Syringothyris chemungensis.* The Millers Hollow member of the Salamanca can be differentiated by representatives of the genus *Pararca* and a characteristic mutant of *"Spirifer" disjunctus.* The occurrence of *Syringospira alta* indicates the presence of the Amity shale fauna. The Chagrin is terminated by the Riceville-Oswayo shale. The black shale equivalent of the Chagrin according to Prosser (1912, p. 515) is the Huron shale.

Charlevoix limestone (Cooper and Warthin).—Occurs on the shore of Lake Michigan in Emmet and Charlevoix counties between the "Upper blue shale" (zone 6) of the Gravel Point limestone and the base of the Petoskey limestone. The "Upper blue shale" is correlated with the upper Alpena (approximately Dock Street clay) of Alpena County. The lower part of the Petoskey limestone at Bayshore is correlated with the Norway Point formation of Alpena County. Thus the Charlevoix occupies the position of the Four Mile Dam limestone and consists of a heterogeneous series of limestones ranging from platy lithographic limestone with green shale partings to limestone with coral and crinoidal debris and oölites. These all suggest near-shore action and accumulation. It is therefore suggested that the Charlevoix may represent a western shore phase of the Four Mile Dam limestone, but no fossils yet discovered support this contention.

Chattanooga shale.—At its type section Swartz (1924) determined the Chattanooga shale to be in part Upper Devonian and in part Lower Mississippian. Here the section consists of about 11 feet of shale: The lowest 8 feet corresponds to the Cleveland shale; the next overlying 9 inches of mottled shale are assigned to the Bedford-Berea; the next ⅜ inch of black shale is said to be of Sunbury age; over the Sunbury 2 feet 4 inches of hard shale (Glendale) is correlated with the lower Cuyahoga of Ohio. These are long-ranging correlations, to say the least, and are not based on the best of evidence, but if Swartz is correct the main body of the Chattanooga at its type locality is correlated with the Cleveland shale which is here regarded as Devonian. Correlation of these black shales is so difficult that the age of each occurrence must be considered on its own merits. In northeastern Arkansas black shales called Chattanooga contain Middle Devonian or lowest Upper Devonian fossils; these occurrences should be renamed or referred to the Trousdale shale, Geneseo equivalent in West Tennessee.

Clam Bank series.—Schuchert and Dunbar (1934, p. 104-106) describe a series of red clastics exposed along Clam Bank Cove, St. George Peninsula, Newfoundland. These contain a few fossils, among them a Spirifer "that is more like a Devonian than a Silurian form" and *Camarocrinus,* the bulbous holdfast of *Scyphocrinus.* The evidence is certainly not conclusive of a Devonian age for these sediments because *Camarocrinus* is as well known a Silurian form as it is a Devonian one, ranging from the Keyser to New Scotland. The Clam Bank series may be Devonian but it may also be equivalent to the Keyser which is now regarded as Silurian (Swartz, 1939). It may also be equivalent to the *Camarocrinus*-bearing Griffon Cove River formation of Kindle (*which see*) exposed on the north side of the Gaspe Peninsula.

Cleveland shale.—Dark-gray and black shale of probable Devonian age. Underlies the Bedford shale which contains the Hamburg-Glen Park fauna and overlies the Huron shale, probable black shale facies of the Chagrin shale of Conewango age. The Cleveland shale thickens westward. According to Caster (1934, p. 151) its Devonian age is shown by its fossils, and its eastward thinning represents a reversal of the usual east-west encroachment. Caster suggests uplift of the Cincinnati mass as the source of the black shale, the black muds being poured into the sea from the west producing a wedge thinning eastward. As the Cleveland underlies the Bedford its designation as probably Devonian is strengthened.

Clifty limestone.—Purdue and Miser (1916, p. 9) list a few Hamilton species, among them *Tropidoleptus carinatus and "Spirifer" sculptilis.* In addition to the listed forms the writer identified *Vitulina pustulosa* and *Centronella impressa* in the same collection (collection in U. S. National Museum, Cooper locality no. 401l. These four fossils indicate a correlation with the *Vitulina* zone of the Lingle limestone of Illinois.

Cooper limestone.—Correlation with the Coggon limestone is here suggested. Both contain a small *Emanuella* suggestive of *E. subumbona* of the New York Hamilton and Tully. A large snail, *Turbonopsis providencis* (Brodhead), occurs in the Cooper and is similar to a species in a lithographic limestone at the top of the Potter Farm formation of Michigan. Therefore the Cooper is tentatively placed with the Cedar Valley in the Tully, and equivalence with the Potter Farm is indicated.

Cornwall shale.—Willard (1937, p. 271-272) assigns this shale to the Onondaga, but it may include Bakoven as well (Warthin, personal communication).

Darby limestone.—Doubtful if any Threeforks is present.

Delaware limestone (Cooper and Warthin).—Covers a greater area than hitherto believed At its type section several zones can be recognized, one of the most important, the *Hadrophyllum d'orbignyi* zone, near the top. This zone can be recognized at the top of the Speeds limestone in Clark County, Indiana. Peoples (1931, p 435-436) reports this fossil from the Pegram limestone of western central Tennessee.

The upper Columbus limestone of northwestern Ohio and the Dundee of southern Michigan containing *Spirifer lucasensis* Stauffer are regarded as of Delaware age The lower part of the Dundee of Presque Isle County, Michigan, contains *S lucasensis* and is assigned to the Delaware.

The coral-reef limestone of Formosa, Ontario, and vicinity has been correlated erroneously with the Onondaga (Pohl, 1930, p. 59), but as it overlies Detroit River dolomite containing "*Spirifer*" *divaricatus* it is regarded as of Delaware age.

Delphi black shale.—Assigned to the Naples and Chemung because of the presence of three species of *Manticoceras*. Two of these species according to Miller (1938, p. 83, 91, 123) resemble Genundewa forms, while the third, *M. unduloconstrictum,* is more like a Chemung species.

Detroit River group (A. S. Warthin).—As now understood includes the Sylvania sandstone as its basal deposit (Carman, 1936, p. 260). Reaches its maximum thickness of 1000 feet in wells in central part of southern peninsula of Michigan, but the outcrop thickness is usually less than 200 feet.

Wherever the base of the Sylvania is exposed its contact with the underlying rock is disconformable and the disconformity widens on leaving the Michigan basin. In western Lake Erie region the Sylvania lies on the Bass Island (Upper Silurian) beds, but the unconformity includes at least part of the Lower Devonian, as indicated by Lower Devonian ostracoderms in a pre-Sylvania stream channel. If this unconformity is followed to the farthest limits of the Michigan basin province, the base of the "Sylvania" basal sand will of course rise in the column. This far-out "Sylvania" sand is probably to be found in the Pendleton of Indiana, the Hillsboro sandstone of southern Ohio, and the Springvale sandstone in the region around Cayuga, Ontario. In the first two cases the unconformity beneath the sand is so wide that it rests upon Niagaran limestones. Being younger than the true Sylvania, the Pendleton is overlain by higher (Middle Devonian) beds; the Hillsboro is overlain by the Ohio shale.

The beds covering the Detroit River vary somewhat in age from place to place indicating a disconformity at the top, but not so great as the one at the base. The overlying rocks at Sibley, Michigan, northwestern Ohio, and Amherstburg, Ontario, are of Dundee age; in Bruce and Huron counties, Ontario, the overlying rock is lower Onondaga with *Amphigenia,* except for the reef masses at Formosa and vicinity which are of Delaware age; on Mill Creek, Cheboygan County, Michigan, the Detroit River is said to contain a Schoharie fauna and is overlain by the *Amphigenia*-bearing Mackinac breccia.

Sherzer and Grabau (1910) subdivided the Detroit River in descending order into the Lucas dolomite, Amherstburg dolomite, Anderdon limestone, and Flat Rock dolomite. This succession nowhere crops out in its entirety. In the Oakwood salt shaft at Detroit the Amherstburg is not definitely recognized, although in the Detroit River channel cuts, only 3½ miles away, a thick section of both beds occurs. Two miles east of the channel cuts, at Amherstburg, Ontario, the Lucas and Amherstburg again cannot be recognized. This has led to the belief that the three upper subdivisions of the Detroit River group have either very local and scattered distribution or they are facies of a single formation. The Anderdon limestone may be only a limestone facies of the Lucas (Ehlers, personal communication), and the Amherstburg a dolomitized coralline facies of the same formation. If the Amherstburg and Anderdon are considered facies of the Lucas dolomite no grounds can be found for recognizing the Flat Rock as anything but the lower part of the Lucas beds. Furthermore, part of the Detroit River in Ontario and Ohio may represent dolomitized Onondaga and Columbus limestones.

Devils Gate limestone.—Embraces 2065 feet of limestone lying on the *Stringocephalus* zone of the Nevada limestone, Eureka district, Nevada. The type section is at Devils Gate, 6 miles west-northwest of Eureka.

According to Merriam (1940, p. 16, 25, 59-61) this limestone contains four zones: the lowest, *Stromoatopora* zone, the *Spirifer argentarius* zone, the *Pachyphyllum* zone, and finally the *Cyrtospirifer* zone. The age of the lowest zone cannot be definitely stated, but it is tentatively assigned to the Genesee or lowest group of the Upper Devonian. The *Spirifer argentarius* zone contains *Tenticospirifer utahensis,* a multiplicate spiriferoid of late Devonian type possibly related to *"S." cyrtiniformis* and very close to *"Trigonotreta" shellrockensis* Belanski. On the basis of the spiriferoids the zone is placed low in the Naples group. The *Pachyphyllum* zone contains a species of *Hypothyridina* and *Cyrtospirifer,* a combination relating it to the Independence fauna and the Chemung. The uppermost or *Cyrtospirifer* zone containing *Athyris angelicoides* Merriam is assigned to the Cassadaga (Canadaway group) stage because of its similarity to the New York fauna.

Dock Street clay.—See Alpena limestone.

Dundee limestone.—Bassett (1935, p. 425-462) shows that the type Dundee exposed in the Sibley Quarry, southeastern Michigan, contains *Spirifer lucasensis,* and is underlain by the Anderdon limestone. He correlates this section with the "Upper Columbus" of the Silica and Whitehouse quarries in northwestern Ohio where the Dundee overlies Lucas dolomite and underlies the "Blue bed" of the Silica shale. Thus the Dundee lies between Onondaga dolomite and Hamilton shale. In Presque Isle County, Michigan, the Dundee overlies the Mackinac Breccia of lower Onondaga (Schoharie) age and underlies the Rogers City limestone which carries the *Stringocephalus* fauna.

Dutch Creek sandstone.—Weller's (1939, p. 130) recently suggested correlation of the Dutch Creek and Beauvais sandstones on the basis of lithology is regarded as wholly untenable. *Amphigenia* is abundant in the Dutch Creek but has never been seen in the Beauvais. Moreover, *Rensselandia* is abundant in part of the Beauvais but is unknown from the Dutch Creek. In all known occurrences *Amphigenia* is lower Onondaga, and *Rensselandia* is Hamilton to Tully.

Ellsworth shale (G. M. Ehlers).—Has been placed in the Mississippian because the upper part of the underlying Antrim shale was assigned to the Mississippian by Doctor Ulrich.

Available information suggests that the upper as well as the lower part of the Antrim is Upper Devonian. Dr. Margaret Morse Rainwater, as the result of a study of the conodonts of the Antrim, which soon will be ready for publication, concluded that the conodonts of the upper part of this formation were more closely related to Devonian species than to species from strata considered of undisputed Mississippian age. The identification of a few cephalopods obtained from the upper part of the Antrim also indicates a Devonian age. Professor A. K. Miller, who identified the cephalopods for Doctor Rainwater as *Tornoceras* sp., states that *Tornoceras* never has been found above the Devonian and not even in the highest Devonian of this country and Europe.

If the Antrim is Upper Devonian, the Ellsworth shale also may be of this age. No one, so far as I know, has collected any fossils from the Ellsworth which definitely prove its geologic age. An abundance of *Sporangites* in the Ellsworth and Antrim shales and an interfingering of these shales as indicated from the examination of well samples by petroleum geologists suggest, however, that the two formations do not differ greatly in age. From the evidence available I believe the Ellsworth and Antrim shales should be placed in the Upper Devonian rather than in the Mississippian.

The relationship of the Ellsworth shale, which is present in the western half of the Southern Peninsula, to the so-called Bedford, Berea, and Sunbury formations of eastern Michigan is not settled. Records of deep wells in eastern Michigan given by petroleum and other geologists indicate that the "Bedford" rests on the Antrim shale and that the Coldwater shale lies on the "Sunbury shale" or "Berea sandstone" where the former is absent. Well records in western Michigan show the Coldwater directly on the Ellsworth shale. The fact that the Coldwater shale rests on different strata in passing from the eastern to the western part of the Southern Peninsula suggests that a systemic boundary may be present at the base of the Coldwater. The finding and identification of fossils in cuttings from deep wells would aid materially in the recognition of the boundary, the presence or absence of an overlap, and especially the stratigraphic relationship of the Ellsworth.

I have never admitted that the "Bedford", "Berea" and "Sunbury" of Michigan are the equivalent of the formations in Ohio having these names. The Michigan formations are on the western side of the Cincinnati geanticline in a different basin of deposition from that of the typical Ohio formations which are on the eastern side of this structure about 200 miles distant. I think it is quite possible the "Bedford", "Berea", and "Sunbury" of Michigan may be Devonian. A comparison of stratigraphic positions (which may be apparent rather than real) and of thickness indicates that these formations are equivalent to some part of the Ellsworth which I should place in the Devonian.

The Coldwater shale and the succeeding Marshall Group undoubtedly are Mississippian. The Coldwater is the first, apparently the lowest, formation, we can say is definitely Mississippian age. Some distance up in the Marshall occurs a fauna with many goniatites identical with species found in the Rockford limestone of Indiana. The stratigraphic tie-up thus indicated means that we have a very large Kinderhook section in Michigan, most of which, unfortunately, is covered by drift.

Elm Point limestone.—Buff limestone about 20 feet thick containing a small fauna which Kindle (1914, p. 251) listed, including a fine-ribbed *Atrypa* with prominent marginal frill and the snail *Büchelia tyrrelli* (Whiteaves). This fine-ribbed, frilled *Atrypa* is common in the Rogers City limestone of Michigan and occurs with the same snail. The presence of *Büchelia tyrrelli* in the Manitoba limestone as well as the Elm Point shows that all three limestones of Manitoba—the Elm Point, Winnepegosis, and Manitoba—are not to be dissociated.

Enfield formation.—As represented on the chart this formation lies between the Ithaca and Chemung and, according to Chadwick (1935, p. 858), is the eastern representative of the Rhinestreet-Hatch sequence. A correlation varying considerably from that on the chart was published in abstract form by Caster (1933, p. 201, 202) who claims that the Enfield of the Ithaca region consists of the Van Etten and Hatch, the former "presumably" of Wiscoy age and including the First *Tropidoleptus* zone at its base Inasmuch as the Wiscoy is generally represented as overlying the Nunda sandstone (High Point) high in the Chemung, Caster's view is in serious conflict with Chadwick's

Escuminac formation (William L. Bryant).—The following information was furnished by Bryant after request by Cooper to analyze the list of Escuminac fishes given by Alcock (1935, p. 88).

> "The genera *Fleurantia* and *Euphanerops* are known only from Scaumenac Bay. *Scaumenacia* is also known from the Upper Old Red Sandstone *Bothriolepis* is a typically Upper Devonian fish, as is *Eusthenopteron* (also found in Europe). *Cephalaspis* ranges from the Upper Silurian to the Upper Devonian, *Cheirolepis* from the Middle to the Upper Devonian; *Coccosteus* mostly from the Upper Devonian though it is also known from the Middle Devonian. *Diplicanthus* is found from the Upper Silurian to the Upper Devonian. *Holoptychius* occurs both in the Middle and Upper Devonian. The *Acanthodians* are known from the Lower Devonian to the Permian Westoll and Graham-Smith were here last year and . . In a general way they consider the deposits of Scaumenac Bay to be of *Early* Upper Devonian Age, and in that I agree"

Esopus shale.—Hitherto classified as the upper member of the Oriskany, has been transferred by Willard (1939, p. 145) to the Onesquethaw stage because of transition between the two in eastern New York and Pennsylvania. The upward transition of the Esopus into the overlying formation is not with the Onondaga but with the Schoharie which has been mistaken for Onondaga. The fauna reported as belonging to the Esopus is actually a calcareous-shaly facies of the Schoharie as shown below by Goldring and Flower. The upper transition zone links the Esopus to the Schoharie-Onondaga, and it has been so placed on the chart.

Exshaw shale.—In the Canadian Rockies, consists of dark shale between the Mississippian Banff limestone and the Minnewanka limestone containing *Cyrtospirifer.* Contains *Camarotoechia nordeggi, Cyrtospirifer "whitneyi", "Spirifer" raymondi, Athyris "angelica"*, and *Tornoceras* (Warren, 1937, p. 454). *Tornoceras* indicates the Devonian age of the formation, and the other species suggest correlation with the Threeforks shale and the shaly part of the Milligen formation of Idaho.

Famine series.—Located east and northeast of the junction of the Famine and Chaudiere rivers south-southeast of Quebec. Consists of a basal conglomerate, intermediate shaly limestone, and an overlying shale, all occupying a belt about 1500 feet wide. The limestone contains fossils including the following: *Leptocoelia flabellites, Spirifer gregarius* [?], *Chonetes arcuata, Stropheodonta, Pterinea textilis*, and *Paracyclas*, which suggest a correlation with the upper Grande Grêve limestone.

Fourmile Brook shale.—Is about 600 feet thick and was named by Kindle (1938, p. 39) to distinguish the upper part of Alcock's Heppel sandstone. Kindle correlates this shale with the Ithaca of New York primarily on the presence of the brachiopod *Spirifer pennatus posterus* Hall and Clarke, which, in New York and the Appalachians, is common in the lower part of the sandy facies of Taghanic and Naples groups. Contributory evidence, according to Kindle, is furnished by the presence of *Strophalosia truncata* var. and *Paracyclas lirata.*

Other listed fossils do not support an Ithaca age for this shale. For example, the Lower to Middle Devonian genus *Eodevonaria* is listed. In fact none of the genera recorded is of unequivocal Upper Devonian age. Kindle furnishes no figures of the *Spirifer pennatus posterus;* consequently the correctness of the identification cannot be confirmed. It is an important fact that *Eodevonaria* has not been reported above the Onondaga. The listed species are most suggestive of a Hamilton age, as they are given Hamilton names, but *Eodevonaria* argues for an earlier date. The shaly facies of the Onondaga in Pennsylvania contains many Hamilton species; therefore the Fourmile Brook shale fauna is probably actually an Onondaga fauna normally restricted to the shale facies.

Four Mile Dam limestone (Cooper and Warthin, 1942).—At its type section on Thunder Bay River above Alpena consists of reef limestone containing a fauna with Centerfield elements: *Spirifer venustus, Cyclorhina, Camarospira, Camerophoria, Parazyga, Nucleocrinus, Strombodes* *alpenensis*. This fauna is also known from a few feet of gray limestone overlying the Dock Street clay in the southeast side of the quarry of the Thunder Bay Quarries Company, Alpena, Michigan. This limestone furnishes the most definite link between Michigan and Ontario as it correlates with the Hungry Hollow formation.

Gaspe sandstone.—Age hitherto variously stated between the Oriskany and the Hamilton, but recently Kindle (1938, p. 41, 44) suggested an Upper Devonian ("Portage") age for the upper part of the Gaspe (Sonneau Brook beds-Fourmile Brook shale) sandstone. For the lower part of York River sandstone Kindle suggests a Hamilton age, mainly on the basis of Mollusca. Kindle correctly indicates that the Gaspe differs strongly in facies from the underlying Grande Grêve limestone which contains a related fauna, but the writer believes the Gaspe fauna to be an Onondaga fauna—a sandstone facies of the Onondaga.

The known fauna of the Gaspe sandstone comprises 16 brachiopods, 25 pelecypods, 12 gastropods, and about 16 species from other phyla. According to Kindle seven brachiopods, *Eodevonaria gaspensis, Chonostrophia dawsoni, Spirifer gaspensis, Leptocoelia flabellites, Eatonia peculiaris, Rensselaeria gaspensis,* and *Chonetes billingsi,* are holdovers from the Grande Grêve fauna and, as relicts, have no correlative value. Twenty-three species or one-third of the fauna are closely allied to Hamilton forms, mostly long-ranging generalized Mollusca such as *Leda (Nuculana), Paracyclas, Nuculites, Schizodus, Sphenotus, Modiella, Grammysia,* and *"Actinopteria"*. The poor preservation of molluscs in the Gaspe sandstone makes these forms very difficult to identify with known Hamilton species and makes questionable many of the specific determinations recorded. The writer therefore believes that the suggested correlation places the Gaspe fauna too high in the Devonian for the following reasons:

(1) The so-called relict forms are numerically the most abundant in individuals, far outnumbering the supposed advanced types. *Chonostrophia* reaches its greatest size in this formation.

(2) The pelecypods are genera chiefly restricted to sandstone and are found in any arenaceous facies from the high Helderberg through the remainder of the Devonian. They are facies fossils, and therefore their correlative value should be discounted.

(3) The sandy equivalents of the Upper Grande Grêve faunas are not known. In the Appalachians where the Onondaga is shaly or sandy, the same types of clams appear as in the Gaspe sandstone (Kindle, 1912). Furthermore the so-called *Actinopteria* or *Pterinea* of the Gaspe sandstone is a type not yet reported from the Hamilton (group of *P. com-*

munis) but is common in the Helderberg and Onondaga. From the above arguments the writer believes that the Gaspe sandstone is of Schoharie (Camden chert) age.

Geneseo shale.—Assigned to the Taghanic stage. In the Cayuga Lake region thin limestones in the base of the Geneseo indicate a persistence of Tully deposition although the base of the Geneseo to the west is disconformable on the Tully. Some of the most important elements of the Geneseo such as *Schizobolus truncatus* and *Buchiola* appear in the Hamilton. Furthermore the Tully fossil *Hypothyridina venustula* occurs in the black shale of eastern Kentucky hitherto regarded as Geneseo, thus supplying a faunal as well as a lithologic link to the Tully. The writer found *H. venustula* in the collection of fossils made by G. H. Girty at Oil Springs on Lulbegrud Creek, 2 miles southeast of Indian Fields, Clark County, Kentucky. This collection came from the *Lingulipora* zone of the Geneseo and points either to a Tully age for the Geneseo or a greater range of *Hypothyridina venustula.*

The most important reason for assigning the Geneseo to the Tully stage is the postulated presence of a disconformity at the top between Lake Erie and Seneca Lake. Over this distance the Genundewa or *Styliolina* limestone overlies the Geneseo and is composed almost wholly of tiny shells of the pteropod *Styliolina.* Accumulation of this limestone, even though it is only a foot thick, must have taken considerable time and thus represents a distinct break in black shale deposition. Not only does the Genundewa represent a break in prevailing sedimentation, it actually contains many new faunal elements—those of the *Manticoceras* zone. It is in the Genundewa limestone that all but one of the goniatites assigned to the "Genesee" shale appears; it is at this level that the Naples fauna becomes established.

In Kentucky and the Midwest to eastern Arkansas and Texas a representative of the Geneseo is known in the form of black shale containing the brachiopod *Schizobolus.* In Michigan and Iowa the Thunder Bay and Cedar Valley formations are possible correlates of the Geneseo but at present seem best placed with the Tully limestone.

Geneva dolomite.—According to Sutton and Sutton (1937, p. 331)

> "The Geneva [of southern Indiana] is a northward lithologic facies of the Jeffersonville formation of Onondaga age as shown by the Onondaga faunules of the Jeffersonville which continue into the Geneva at the same horizons and by the lateral transition from the dolomitic Geneva limestone to the purer Jeffersonville limestone. This gradation begins farthest south in the lowest beds and reaches the progressively higher beds toward the north, which makes it appear on casual investigation that the Jeffersonville thins toward the north, the thinning being accompanied by a corresponding thickening of the Geneva."

Campbell (1942) strongly dissents from this view and claims that the Geneva is always distinguishable from the Jeffersonville.

Genshaw limestone (Warthin and Cooper).—Originally proposed (Warthin and Cooper, 1935, p. 526) for the thin limestones and shales between the Ferron Point formation and the Killians limestone. Later work, particularly in Presque Isle and Cheboygan counties, showed that black limestone of Killians facies appears in the Genshaw. The writers therefore redefine the Genshaw to include the Killians and the lower 15 feet of the Alpena limestone which contains many Genshaw species.

Genundewa limestone.—Represents a break in sedimentation (*see* Geneseo) along the northern shore of the Devonian sea. This break taken in connection with the influx of many new types of fossils is believed to inaugurate the Upper Devonian in New York. West of New York the Squaw Bay limestone is believed to represent the Genundewa and contains the goniatite *Koenenites,* earliest of the Upper Devonian goniatites of Germany.

Glen Park limestone.—For years regarded as Kinderhook in age and correlated with the Hamburg oölite of Illinois. It is suggested here that the Glen Park fauna also correlates with the Bedford of Ohio and the marine fauna at the top of the New Albany shale. Branson and Mehl (1938, p. 157; Branson, 1938, p. 8, Fig. 2) suggest that the Louisiana limestone and Glen Park faunas are actually Devonian. The Bedford, Louisiana, and Glen Park contain many Devonian elements such as *Atrypa, Nucleospira, Productella, "Delthyris", Cypricardella, Sphenotus,* and *Pholadella.* Most of the reported clams and snails are Devonian types. A few Mississippian types appear such as *Spirifer* (*S. marionensis* type) and "Goniatites". *Syringothyris,* often quoted as indicative of the Mississippian, is also a well-established Devonian fossil in southwestern New York and northwestern Pennsylvania. It is not important here to settle the age of the Glen Park and its correlates; it is significant to emphasize the equivocal nature of the fauna and the possibility of its serving as a base of the Mississippian or top of the Devonian when it is finally decided on which side of the line it should be placed. The fairly widespread character of this fauna enhances its value for the purpose suggested.

Grande Grêve limestone.—Clarke (1908, p. 39-46) divided this limestone into three parts, the lower two containing an abundance of Oriskany fossils, while the upper division contains few of these familiar forms. Some who have studied the Grande Grêve limestone see a similarity to the Schoharie, but the writer believes that the entire sequence is Oriskany. However, in all probability the top division is higher than any Oriskany exposed in New York. The upper division contains such Oriskany forms as *Rensselaeria, Chonostrophia, Eodevonaria, Anoplia, Spirifer arenosus,* and *Eatonia.* Except for the last two these are seen higher in the Gaspe sandstone, and all but *Rensselaeria* and *S. arenosus* are known in the Camden chert. The writer has placed the entire Grande Grêve in the Deerpark stage because no unequivocal Onondaga forms have yet been taken from it, and the higher Gaspe sandstone fauna is believed to represent the Onondaga (Schoharie).

Grand Tower limestone.—Savage (1910) listed the Grand Tower fauna and reports *Paraspirifer acuminatus* and *"Spirifer" gregarius* in the upper part. In the Little Saline Valley the Grand Tower contains a coral bed in the lower part, and the upper layers are crowded with *Schizophoria.* These upper beds may belong to the Marcellus rather than the Onondaga. Beds containing *Spirifer lucasensis* occurring far below *Microcyclus* at the Bakeoven have been assigned to the Marcellus.

Gravel Point limestone (Cooper and Warthin).—At the base of the formation a foot of calcareous shale contains fossils like those in a blue shale 40 feet below the top of the Alpena limestone. At the top of the Gravel Point another blue shale, "the upper blue shale," or zone 6 of Pohl, contains crinoids and brachiopods similar to those of the Dock Street clay at the top of the Alpena. Therefore the interval occupied by the Gravel Point is essentially that of the upper 40 feet of the Alpena limestone. The *Chonetes emmetensis,* so abundant in the lower part of the Gravel Point limestone, also occurs at the base of the upper 40 feet of the Alpena.

Griffon Cove River beds.—Located between Fox River and the Forillon, Gaspe Peninsula, consist of red shale and gray limestone. The crinoid *Scyphocrinus* [*Camarocrinus*] is reported near the top of the lower quarter of the formation, and a small fauna of brachiopods is reported a little higher in the section. Kindle (1938, p. 20) lists *Spirifer vanuxemi* var., *Rhynchospirina,* and *Whitfieldella* and correlates the beds with the Keyser limestone of Maryland which is here regarded as Silurian. This sequence may also be related to the Clam Bank series of Schuchert and Dunbar.

Guilmette formation.—*Stringocephalus* occurs about 100 feet above the base, and *Platyschisma mccoyi* was found near the top. The lower part is therefore correlated with the *Stringocephalus* zone, and the upper part with the *Spirifer argentarius* zone of the Devils Gate formation which contains *Platyschisma.*

Harrogate limestone.—Proposed by Shepard (1926, p. 626) for Devonian beds exposed not far northeast of Harrogate in the Brisco-Dogtooth area of the Kootenay District, British Columbia. Since the naming of the formation Evans (1933, p. 142-145) discovered additional exposures in the same area. Fossils collected are listed by Kindle in Evans' report. A collection in the U. S. National Museum made by Dr. Edwin Kirk from the section east of Harrogate contains *Schizophoria "macfarlani," "Martinia" meristoides,* large *Productella, Atrypa,* and the coral *Mcgeea.* These fossils, particularly the first and last, suggest an Upper Devonian age (Chemung). Kindle on the other hand regards the same assemblage as Middle Devonian and relates it to the fauna of the Pine Point formation occurring under *Stringocephalus* in the Great Slave Lake area.

Heppel formation.—At least 3500 feet of sandstone and shale correlated with the Gaspe sandstone. Overlies the Causapscal formation (Alcock, 1935, p. 83-85).

Hungry Hollow formation (Warthin and Cooper, 1941).—New name used for the "Encrinal" and coral bed of the Widder in the southern peninsula of Ontario. Contains most of the Centerfield fauna.

Huron shale.—This name has been loosely used for black shales exposed in northern Ohio, Ontario, and part of Michigan. It has been regarded as "Portage" by some geologists and younger by others, depending on the interpretation of its relationship to the beds overlying it and to the Chagrin shale. According to Ulrich (1912, p. 165-166) the Huron shale is Mississippian and unconformably overlies the Chagrin shale to the east. In the same direction the Huron is overlapped by the Olmstead and Cleveland shales.

Prosser interprets the Huron shale as a facies of the Chagrin representing the passage of the Chagrin from a gray shale having a "Portage" facies to a black shale west of Cleveland. The writer agrees with this view. The Chagrin has been shown to be the sandy shale facies of the Conewango. It would be expected therefore that, as one traced the Chagrin westward, it would pass laterally into a black shale as most of the other Devonian members do.

A Mississippian age for the Huron shale seems to the writer most unlikely as it is overlain by the Cleveland and the Bedford shale. It is shown elsewhere (*see* Bedford shale) that Girty has dated the fauna of the Bedford shale as Devonian, and that it seems to correlate with the fauna of the Hamburg-Glen Park of the Kinderhook. Without arguing the Devonian or Mississippian age of this fauna, the writer has selected it as the upper limit of the Devonian.

Hypothyridina.—Because the various members of this genus have been widely and often incorrectly used to identify the base of the Upper Devonian, a discussion of this interesting rhynchonellid is important. Its rotund form and square front make the genus easy to recognize, but the identification of species is difficult. The known species are variable, a fact that has led to unfortunate assignments of specimens from widely separated parts of the Devonian column generally to *H. cuboides*. Stratigraphers have also fallen into the habit of regarding most occurrences of *Hypothyridina* as early Upper Devonian. Examination of the age of this genus in North America and Europe shows that this practice is not justified. This note is written in the hope that *Hypothyridina* will be used more cautiously as a guide fossil in the future.

In Europe *Hypothyridina* ranges from the Eifelian (Schulz, 1883) (Nohnerkalk with *H. procuboides*) through the *Stringocephalus* zone or Givetian *(H. procuboides, H. demissa)* (Torley, 1934) into and through the middle Frasnian (*H. cuboides* and varieties) (Maillieux, 1940)—*i.e.*, in Europe the genus ranges from early Middle Devonian through the middle of the Upper Devonian.

In the United States and Canada *Hypothyridina* is known at several levels and is represented by several species. A possible new species of doubtful affinities occurs in the lower Nevada limestone. *Hypothyridina venustula* is abundant in the Tully of New York; it occurs in the same formation in Pennsylvania and is regarded here as Middle Devonian. *H. intermedia* occurs in the Solon member of the Cedar Valley formation of Iowa which is correlated with the Tully. The *H. magister* is characteristic of the Shellrock which is placed on the chart as a calcareous facies of the Naples group. In the Devils Gate limestone of Nevada two species of *Hypothyridina* are known which belong in the Finger Lakes to Chemung stages. The higher one, *H. emmonsi*, occurs in the *Pachyphyllum* zone.

In the Great Slave Lake and Carcajou Mountain regions of northwestern Canada *Hypothyridina* occurs with *Cyrtospirifer disjunctus* at a horizon far above the Naples in the Chemung stage. A species of *Hypothyridina* identified by Stainbrook (1935, p. 254) as *H. emmonsi* occurs in the Independence shale of Iowa in association with a fauna like that of the Canadian occurrences. Thus in America the range of *Hypothyridina* is from high in the Lower Devonian to the middle of the Upper Devonian.

Independence shale (Merrill A. Stainbrook).—The Independence shale lies immediately below the Cedar Valley limestone and above the Davenport member of the Wapsipinicon. In a number of places shale, bearing Independence fossils, may be seen in vertical section below the Cedar Valley and in at least two places above Davenport limestone. In a number of wells, 10 to 20 feet of shale occurs, as shown by the samples, below the Cedar Valley and above the Davenport, and in three of them Independence fossils were recovered from the shale. The discovery pit at Independence penetrated 20 feet of limestone and entered shale at the base of the Cedar Valley. At Fairfax, Iowa, a pit sunk through the basal 5 feet of the Cedar Valley shows several feet of fossiliferous shale immediately below. At no place does the Independence shale occur on Cedar Valley unconformably, and occurrences of the shale abutting Cedar Valley limestone are explainable as due to faulting. The fauna is [of late Devonian age] not the same as that of the Lime Creek [Hackberry] as there are many Independence species not found in that formation and vice versa. Species apparently similar are in some cases specifically or varietally different.

Independence shale (Cooper and Warthin).—The position and age of the Independence shale have long been subjects of debate, one group holding that it normally underlies the Cedar Valley and the other contending that the shale is not in its true stratigraphic position (Fig. 1). The formation was first discovered by Calvin (1878, p. 725) in a shaft sunk below the Cedar Valley at Independence, Iowa. Stainbrook (1935, p. 254) maintains that the shale underlies the Cedar Valley conformably but overlies the Davenport limestone of the Wapsipinicon unconformably and cites as supporting evidence the sinking of an artesian well at Shellsburg which showed Independence shale carrying typical fossils below 65 feet of Cedar Valley limestone.

Despite these contentions purporting to show the Independence below the Cedar Valley, the few known surface exposures and the fauna indicate to the writers that the shale is not in its normal position. In most occurrences shown to the writers by Stainbrook the exposed shale was supported on at least two sides by Cedar Valley limestone. The base of the shale was inferred to rest on Davenport limestone. These exposures suggested sinks filled by shale, and some were so interpreted by Savage (1920, p. 170, 180). The occurrence at Shellsburg, and others like it, are here believed to indicate that the well shafts penetrated caverns and channels long since filled by Independence shale. A similar occurrence of cavern filling is known from Chicago where late Devonian black shale occurs in caverns and crevices in Silurian dolomite (Weller, 1899, p. 483-488). Although the position of the Independence shale is equivocal, its anomalous position is best shown by its fauna.

The fossils of the Independence shale are totally different from those of the Cedar Valley. Whereas the latter contains no unequivocal Upper Devonian species or genera, the Independence contains forms not usually seen below the Upper Devonian in the United States, and not one Independence species is known in the Cedar Valley. The important Independence genera are: *Manticoceras, Ponticeras, Pugnoides, Cyrtospirifer* (related to *C. disjunctus*), *Ambothyris, "Dalmanella" infera* (related to *"D." danbyi), Sulcatostrophia, Douvillinella, Strophonelloides,* and *"Camerophoria" ambigua*. The goniatites (Miller, 1938, p. 71) clearly place this assemblage in the Frasnian of Europe or high in the *Manticoceras* zone of the United States. The presence of *C. disjunctus* and the *"Dalmanella"* place the fauna in the Chemung, the horizon in which the widespread spiriferoid first appears. Similarities are also evident with the Sly Gap shale of eastern New Mexico (San Andres and Sacramento Mountains), the Boule limestone of Alberta, the Carcajou Mountain beds, and Hay River limestone of northwestern Canada. Correlation with the Nunda (High Point) sandstone of New York is also indicated (J. M. Clarke, 1885, p. 72-76). Besides the forms mentioned Stainbrook (1935, p. 254) has identified *Hypothyridina emmonsi* which occurs high in the Devils Gate limestone of Nevada. The Independence fauna is the one assemblage in North America most closely related to the *"Cuboides"* zone of Europe as seen in the Frasnian (F_2) of Belgium.

In summary the writers hold that the Independence shale is not in normal position because: (1) It fills sinks and caverns; (2) its occurrence is spotty, there being no definite outcrop belt; (3) its fauna is that of the Nunda or High Point (Chemung) sandstone and the Frasnian (F_2) of Belgium.

Ipperwash limestone (Cooper and Warthin).—Uppermost formation of the Hamilton exposed along the shore of Lake Huron from Ipperwash Beach to Silica Point and in the vicinity of Ravenswood. Underlies the Kettle Point (Upper Devonian) shale, but correlation with the New York Hamilton is difficult. The Ipperwash contains many fossils, among them *Rhipidomella penelope, Spirifer granulosus, Stropheodonta concava, Tropidoleptus carinatus,* and *Cornellites* [*Pterinea*] *flabellus*. This assemblage indicates a Ludlowville age. Comparison with the Lake Erie sequence shows that the fossils named occur together at two levels only: the basal Wanakah (*Pleurodictyum* zone) and the Tichenor. The writers believe that the Ipperwash fauna is closer to that of the basal Wanakah than to the fauna of the Tichenor. *Pleurodictyum* has not yet been taken from the Ipperwash, but the large *Spirifer granulosus* is most suggestive of the Wanakah. Important Tichenor types such as *Vitulina* and *Centronella* which are abundant on the Lake Erie shore are unknown in the Ipperwash. The evidence stated is neither extensive nor conclusive, and the correlation suggested must be regarded as tentative.

Ironside dolomite.—The fauna is sparse and consists chiefly of corals. In this member in a part of California adjacent to the Goodsprings area north of Sandy Wells road (U. S. Geol. Survey loc. 1994) is a small *"Spirifer"* suggestive of *S. argentarius.* This zone is therefore assigned to the portion of the Devils Gate limestone containing this fossil.

Ithaca formation.—According to Chadwick (1935, p. 858) the Ithaca formation represents the sandy facies of the Middlesex and Cashaqua shales of western New York. Caster (1933, p. 201) in abstract form revised the Ithaca sequence and suggests correlations that vary considerably from Chadwick's views. The following divisions of the Ithaca in descending order are proposed by Caster:

Ithaca formation

Triphammer shale member 140 feet.	(4)
Marathon sandstone member 35 feet	(5)
Williams Brook member 15-25 feet	(6)
Cascadilla shale member 150 feet	(7)
Six Mile shale member 80-100 feet	(8)
Renwick black shale member 8-20 feet	(9)

According to Caster the Triphammer and Marathon members become the Oneonta to the east, and the Six mile shale to the west is equivalent to the Rhinestreet shale overlying the Cashaqua. Thus Caster correlates the Ithaca with a higher formation than does Chadwick. At present it is impossible to settle these important differences; only field work will settle the matter. (*See* Enfield also.)

Jefferson limestone.—Sections in western Montana about Threeforks and in the Phillipsburg Quadrangle show 640-800 feet of gray to black limestone. In the lower 100 feet at Logan, Edwin Kirk (U. S. Geol. Survey loc. 754) collected *"Spirifer" argentarius* and *Tenticospirifer utahensis.* At another locality (no. 673) 2¼ miles northwest of Princeton Kindle (1908, p. 8-10) collected the same two spiriferoids but also *Gypidula* and *Cyrtospirifer* suggestive of *C. whitneyi.* Unfortunately the position of the last two in the section is not known. The listed species are elsewhere known in the lower part of the Devils Gate limestone.

Deiss (1933, p. 41-43) revised the stratigraphy of the Jefferson formation in the Flathead and Lewis and Clark National Forests, separating about 1500 feet of rock into five members in ascending order: White Ridge limestone, Glenn Creek shale, Coopers Lake limestone, Lone Butte limestone, and Spotted Bear limestone. The writer studied fossils collected by Deiss from these limestones and found that the Coopers Lake contained a useful fauna consisting of a few species: *Pachyphyllum, Atrypa* cf. *A. hystrix, "Spirifer"* cf. *S. mesacostalis, Stropheodonta,* and *Schizophoria,* which indicate a Chemung age.

In south-central Idaho, Ross (1934, p. 960) reports the Jefferson in the Bayhorse Quadrangle as 1150 feet of dark dolomite containing abundant corals. In the Borah Peak area, Idaho, the Jefferson contains *Pachyphyllum* among other corals.

In western Wyoming the Jefferson generally consists of buff limestone containing *Atrypa* aff. *A. missouriensis* [1] as its commonest species. In Yellowstone Park Girty (1899) reports a few Jefferson species, among them *Pachyphyllum* and *"Spirifer engelmanni"* which is very close to *Tenticospirifer utahensis.*

The evidence presented suggests that in southwestern Montana and the Yellowstone region the Jefferson should be correlated certainly with the *Spirifer argentarius* zone of the Devils Gate limestone. The presence of *Pachyphyllum* and *Cyrtospirifer* in the same areas indicates that the section probably extends to the *Pachyphyllum* zone of the Devils Gate. In northeastern Utah in Green Canyon, Kindle (1908, p. 16, 17) reports *Spirifer argentarius* and *Cyrtospirifer* from the Jefferson indicating the same range as in Montana and western Wyoming. To the northwest in the Flathead and Lewis and Clark National Forest areas the fauna of the Coopers Lake member suggests a Chemung age.

Kanouse shale.—According to Willard (1937, p. 272) the contained fossils are those of the Esopus (Schoharie?) rather than the Onondaga.

Kennett formation.—The presence of *Calceola* in this formation is not unequivocal evidence of Lower Middle Devonian age, as this coral is known from the Silurian elsewhere in the West. Furthermore, fossils from Gazel, California, collected by C. W. Merriam from this formation and now in the National Museum include *Conchidium* and *Encrinurus* in an assemblage that otherwise suggests Devonian.

Kettle Point black shale.—As pointed out by Kindle and Miller (1939, p. 154) this name has long been overlooked. It was proposed by Logan (1863, p. 387) for the black shales at Kettle Point, Lake Huron, Ontario.

Killians limestone.—*See* Genshaw.

Lake Church formation.—As listed by Cleland (1911, p. 12-20) the Lake Church fauna includes a few important species as follows: a costellate *Atrypa* (Pl. 13, fig. 12) known elsewhere in the Manitoba and Rogers City limestones, *Gypidula "comis"* like those of the Rogers City, *Spirifer* sp. (p. 82, Pl. 14, figs. 13, 14) related to *"S." grieri* and *"S." divaricatus* from the Dundee limestone, and *Trochonema* occurring also in the Rogers City limestone. It is not yet possible to place the Lake Church accurately in the Devonian column, but the fauna as listed suggests correlation with the Dundee and Rogers City limestones of Presque Isle County, Michigan. According to Raasch (1935, p. 261-262) the Belgium or basal member of the Lake Church is duplicated in fauna and lithology by the basal Devonian at Spring Valley, Minnesota. The writer has indicated a correlation of the Lake Church with the *Stringocephalus* zone (Rogers City limestone), and it is therefore interesting to note that Schuchert (1897, p. 417; Stauffer, 1922, p. 408) reported the discovery of loose *Stringocephalus* in Minnesota from a borrowpit just under the Lake Church equivalent.

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[1] This species is not *A. missouriensis* Miller. For a description of true *A. missouriensis, see* Greger (1936).

Lingle limestone.—The middle part of the formation at its type locality contains *Tropidoleptus carinatus, "Spirifer" sculptilis, Vitulina pustulosa,* and *Centronella impressa* which link the formation to the Beechwood of Indiana and Kentucky. Use of the name Lingle for rocks overlying the Grand Tower limestone at Grand Tower is incorrect; these rocks are pre-Lingle (*Vitulina* Zone) and are referred to the St. Laurent formation. (*See* St. Laurent.)

Little Rock Creek limestone (Cooper, 1941).—Occurs just above the Logansport limestone on Little Rock Creek about a mile above Lockport, Indiana, and consists of about 7 feet of brittle, gray limestone containing *"Chonetes manitobensis"*, *"Martinia subumbona"*, *Leptostrophia,* and a few other fossils, all Hamilton types. These fossils may range in age from the Delaware to the Tully. Lithologically the rock is suggestive of the Tully, and this limestone is tentatively correlated with that formation.

Logansport limestone (Cooper and Warthin, 1941).—"Reef" or shelf facies limestone exposed along the Wabash River and its tributaries in the vicinity of Peru and Logansport, Indiana. Contains the same fauna as the Four Mile Dam limestone of Alpena County, Michigan, and the Beechwood of Indiana and Kentucky.

Long Rapids shale.—Black shale containing the brachiopod *"Pugnax pugnus"* and the spores of *Sporangites* (Kindle, 1924, p. 35) exposed in the Moose River region, Ontario. The formation is assigned to the Naples on the basis of the brachiopod.

Mackinac limestone.—Contains the lower Onondaga (Schoharie?) fauna with *Amphigenia elongata.*

Manitoba limestone.—Grayish-brown friable limestone about 80 feet thick at Point Wilkins, 4 miles north of Steeprock River, western Manitoba; overlies the Winnepegosis limestone containing *Stringocephalus.* The fossils listed include the gastropods *Büchelia tyrrelli* and *Omphalocirrus manitobensis* (Kindle, 1914, p. 254-257). The former also occurs in the Elm Point limestone below the Winnepegosis formation, and both species occur together in the Rogers City limestone of Presque Isle County, Michigan. The Manitoba limestone has generally been correlated with the late Devonion *Hypothyridina* zone, but that diagnostic genus has never been found in it. None of the species listed by Kindle from the Manitoba limestone is a late Devonian fossil, and the majority of the genera are those of the Middle Devonian. Therefore, because of the presence of *Büchelia* and the Middle Devonian aspect of the fauna the writer places the Manitoba limestone in the *Stringocephalus* zone, making the Elm Point, Winnepegosis, and Manitoba limestones a unit.

Martin limestone.—Long recognized as a correlate of the Hackberry of Iowa because of abundance of *"Spirifer" hungerfordi, "Spirifer" orestes, Strophonelloides reversus, "Spirifer" cyrtiniformis,* and *Pachyphyllum woodmani.* The presence also of *"Spirifer" mesacostalis* links the formation to the Chemung of the Appalachians and New York.

Milligen formation (Kirk and Cooper).—In the Borah Peak Quadrangle, south-central Idaho, the lower part of the Milligen is shaly and contains species of the Threeforks formation. In a ravine southwest of Freighter Spring, on the east side of Double Spring Canyon (U. S. Geol. Survey loc. 2745) Cooper identified the following species: *Cyrtospirifer monticola* (Haynes), *Camarotoechia* cf. *C. nordeggi* Kindle, globular *Athyris, and Cleiothyridina devonica* Raymond. Other Idaho localities have yielded similar species.

Mineola limestone.—Has yielded a fairly large fauna of generalized and long-ranging Middle Devonian types: *Megistocrinus, Cranaena, Pentamerella* (reported as *Gypidula*), *Pholidostrophia,* and *Conocardium.* The original assignment of the formation to the Onondaga (Branson, 1924, p. 15-24) was based on identification of such species as: *Pentamerella arata, Eunella lincklaeni, Athyris vittata, Schizophoria propinqua,* and the blastoid *Nucleocrinus verneuili,* the absence of dominant Hamilton species, and unconformable relations to the Callaway which was identified as Hamilton. The few Onondaga species listed from the Mineola are, without doubt, incorrectly identified. Of these erroneously identified forms the Mineola *Pentamerella arata* is totally unlike that species, as a comparison with Figures 17-23 on Plate 17 will show. The *Nucleocrinus* according to Branson "are never well preserved. None of them retains the surface markings, and identifications must be made largely on the basis of shape". This genus occurs throughout the Middle Devonian, and species are known from the Hamilton that are difficult to separate from Onondaga forms. Consequently a correlation based on such an identification is insecure.

On the chart the Callaway has been assigned to the Tully, and the Mineola is also assigned there not on the basis of specific identities but because of the Cedar Valley affinities of the fauna. It is important in connection with the Mineola to mention that some outcrops assigned to this formation near Rensselaer in Ralls County are undoubtedly a continuation of the Cedar Valley (Upper Rapid) of Iowa and indicate a close relationship to the Callaway.

Minnewanka limestone.—Forms the bulk of the Devonian of the Rocky Mountains of Alberta. In general contains two faunas: the lower one typified by *"Spirifer" jasperensis* and the upper one by *"Spirifer whitneyi"*. The lower fauna is correlated by Warren (*see* chart) with the Jefferson limestone, while the *"whitneyi"* fauna is placed in the Chemung.

Montebello sandstone—Great local development of sandstone in the Mahantango of Pennsylvania in Perry, Dauphin, and Northumberland counties. Ranges in age from Skaneateles through Ludlowville and possibly into the Moscow. At some localities contains the brachiopod *Rensselandia* in abundance.

Moose River sandstone.—Hitherto correlated with the Oriskany, but certain fossils occurring in it indicate that the age should be extended to the lower Onondaga (Esopus and Schoharie). The presence of *Spirifer arenosus* is sure indication of Oriskany age, while the occurrence of *Amphigenia* (Clarke, 1907, p. 247) indicates the Onesquethaw stage.

Mottville member.—A well-defined sandstone and shell breccia traceable from Cayuga Lake to Susquehanna Valley, New York. From the latter place eastward to Schoharie Valley the Mottville is marked by an abundance of the crinoid *Ctenocrinus*. At Cooperstown and Richmondville, New York, *Paraspirifer* occurs at the Mottville horizon. West of Cayuga Lake the Mottville is believed to be continuous with the Stafford limestone previously placed at the base of the Skaneateles. Because of the presence of *Paraspirifer* in the Mottville and its abundance below the Mottville in eastern New York, it seems best to transfer the Mottville-Stafford to the Marcellus.

Mount Marion sandstone.—*See* Silica shale.

Mowitza shale.—Contains a small fauna having as its most conspicuous element wide-hinged *Cyrtospirifer disjunctus* unlike any other member of the genus yet reported from the West but suggesting types from the New York Chemung.

Muddy Peak limestone.—About 300 feet below recognized Mississippian rocks occurs a fossiliferous zone about 50 feet thick containing corals, *Atrypa*, the snail *Platyschisma mccoyi*, and the brachiopod *Tenticospirifer utahensis*. These fossils indicate equivalence to the *"Spirifer" argentarius* zone of the Devils Gate limestone. In the Goodsprings area not far west of the Muddy Mountains the same fossils appear in the Ironside and Valentine members of the Sultan formation. Thus the Muddy Peak is equivalent to the Ironside and at least part of the Valentine.

Nevada limestone.—As restricted by Merriam (1940, p. 14) contains three distinct faunas in two different types of lithology. The lower or shaly-weathering limestone contains Oriskany and Onondaga types. In the lower part of this division large *Meristella*, *Gypidula (coeymanensis* type), *Acrospirifer (murchisoni* type), and a large *"Spirifer"* suggestive of *S. arenosus* indicate the Oriskany affinities.

Above the Oriskany part of the shaly lower Nevada is a fauna containing *Leptocoelia*, *Acrospirifer*, *Chonetes macrostriata*, large *Dalmanites*, and large *Gypidula*, but most significant is the presence of a large *Calymene* related to *C. platys* of the Schoharie sandstone and lower Onondaga of New York. This species suggests an upper limit for the shaly Nevada which is not higher than lower Onondaga (Schoharie).

The upper part of the Nevada is more massive and often dolomitic; this portion contains two faunas, one suggestive of the Delaware and the other containing *Stringocephalus*. In the lower part (Marcellus equivalent) of the upper Nevada, *Leiorhynchus* and *Styliolina* occur with *Martinia*-like brachiopods strongly suggestive of *"M." maia* (Billings). Small *Pentamerella* collected in this zone on Lone Mountain by Cooper and Bridge suggest a species occurring in the Cherry Valley limestone.

In the dolomitic part of the upper Nevada limestone *Stringocephalus* is abundant in places and is assigned to a position correlative with the fauna of the Winnepegosis limestone. As no recognizable species occur with *Stringocephalus* in the Nevada limestone no alternative is possible.

In the Muddy Mountain area of southern Nevada Longwell found the *Chonetes macrostriata* fauna and *Stringocephalus*, indicating that the Nevada limestone is well represented in that area.

New Albany shale.—It is generally agreed that the lower part of this formation correlates with the Geneseo shale or Tully limestone of New York as it contains *Schizobolus*, *Styliolina*, and *Leiorhynchus quadri-*

costatum. This part of the New Albany is therefore placed in the Taghanic stage at the top of the Middle Devonian.

Above the Geneseo portion occurs about 80 feet of dark shales terminating in a green shale and zone of small concretions containing a marine invertebrate fauna. Above this zone occurs about 5 feet of dark fissile shale; then a disconformity is followed by a foot of shale that is succeeded by the Rockford limestone. The 5 feet of shale contains conodonts of definite Mississippian affinities (Huddle, 1934, Fig. 2). The marine fauna (Huddle, 1933) consists of 10 species that can be correlated with the Bedford shale and Hamburg oölite faunas. The presence of this fauna of latest Upper Devonian or earliest Mississippian age makes it clear that the bulk of the New Albany is to be assigned to the Devonian, as Huddle has already concluded, rather than to the Mississippian.

New Scotland.—In the vicinity of Evans Ferry, Tennessee, Miss Elizabeth Didcoct of Mount Holyoke College collected fossils of New Scotland age, a discovery new for eastern Tennessee.

Newton Creek limestone.—*See* Alpena limestone.

Nunda sandstone (High Point sandstone).—Usually correlated with the Hackberry of Iowa, but the presence of *"Dalmanella" infera, Douvillinella,* and a few other fossils indicates correlation with the Independence rather than the Hackberry as noted by Clarke (1885, p. 72-76).

Olentangy shale.—Dr. John Wells (oral communication) of The Ohio State University reports the presence of the Tully coral *Lopholasma* at the base of this shale indicating at least a partial representation of the Tully.

Onondaga limestone.—In western New York two faunas can be recognized: The lower one having many Schoharie and Camden elements and characterized by *Amphigenia* and many corals is best developed west of New York on the southern peninsula of Ontario. The Schoharie elements include *Centronella, Strophonella ampla, Cyrtinaella biplicata, Calymene platys,* and *Terataspis.* Camden elements are *Eodevonaria* and *Anoplia.* Overlying the *Amphigenia* fauna is one characterized by *Paraspirifer acuminatus* occurring at the top of the Onondaga in New York. It also characterizes the top of the Columbus and Jeffersonville limestones to the southwest but has not yet been found in the Ontario peninsula. So far as known its northernmost range is Pelée Island in Lake Erie and Wayne County, Michigan.

In central New York a third division of the Onondaga, the Seneca limestone, is a dark shaly rock underlying the Marcellus and containing an abundance of *Chonetes lineatus, Dalmanites selenurus,* and large coiled cephalopods. This name had long fallen into disuse, but as the interval it defines is a well-marked unit it is here revived.

Orcas group.—Part of this group is dated as Devonian by an identification of *Atrypa reticularis* reported by Schuchert. These seem to be Middle or Upper Devonian types rather than Silurian forms (McLellan, 1924, p. 217-222).

Paraspirifer.—In Germany *Paraspirifer cultrijugatus* marks the top of the Lower Devonian and the base of the Middle Devonian, but in Belgium the *Paraspirifer* zone is regarded as the base of the Middle Devonian.

In the United States *Paraspirifer* is abundant at the top of the Onondaga, Columbus, and Jeffersonville limestones and is reported from the base of the Onondaga in Ontario. In New York the genus ranges out of the Onondaga for more than 1000 feet into the Mottville member at the top of the Marcellus. It is also a prominent member of the Silica shale fauna. *Paraspirifer* thus characterizes the Marcellus fauna as well as that of the Onondaga. In comparison with the European sequence on the basis of the range of *Paraspirifer,* the Onondaga appears to correlate with the Upper Coblenzian rather than with the Eifelian as hitherto claimed. The Marcellus then would correlate with the Eifelian or upper *Paraspirifer* zone. In this connection it is interesting to note that the *Stringocephalus* limestone (Rogers City) rests on the Marcellus (Dundee) limestone rather than the Onondaga.

Pegram limestone.—Recently this formation was examined in detail by Peoples (1931, p. 431-439) who suggests that the name Pegram be dropped in favor of Jeffersonville for the Onondaga part and Sellersburg for the Hamilton portion. Peoples gives a long list of fossils from the Pegram of the Harpeth River Valley, west of Nashville, Tennessee, which shows the presence of many Hamilton species but few of Onondaga affinities. Of the latter *Nucleocrinus verneuili* is often misidentified, and blastoids of this type are now well known in the Hamilton. The writer examined the specimens of *"Spirifer" gregarius* reported by Peoples and found one of them to be an *Elytha* and the other a spiriferoid of unknown affinities. Doubt is therefore cast on the unequivocal Onondaga species reported.

After examination of the Pegram exposures on Mill Creek in the Waynesboro Quadrangle and the exposures on U. S. highway 70 in the Harpeth Valley the writer concluded that the part of the Pegram containing *Tropidoleptus* and *Hadrophyllum* is definitely of Hamilton age but that the lower part may be Hamilton or Onondaga; at all exposures seen no fossils of unequivocal Onondaga age were collected.

Besides the areas of Pegram mentioned Dunbar (1919, p. 101) reports this limestone over calcareous Camden at the "Whirl" north of Bakerville, Tennessee. This section exposes the only known calcareous Camden, and over a covered interval of 18 inches occurs a light-gray limestone composed almost wholly of corals both simple and compound that is assigned to the Pegram. Inasmuch as the Camden contains corals and this area is some distance from the main area of Pegram outcrop it is suggested that the bed probably belongs to the Camden rather than to the Pegram.

Pendleton sandstone.—Study of a large collection of Pendleton fossils made by Dr. John M. Clarke reveals a relationship of the Pendleton fauna with fossils of the Amherstburg dolomite. The peculiar and little known spiriferoid *Prosserella subtransversa* occurring with specimens of *Panenka, Pentamerella* aff. *P. arata, Proetus crassimarginatus,* and abundance of a small *Atrypa* (Sherzer and Grabau, 1910, p. 162, Pl. 20, fig. 1) indicate the suggested correlation. The Amherstburg dolomite has been correlated with the Lower Onondaga which contains many Schoharie elements. Inasmuch as the Lucas and Amherstburg may be facies of one formation the Pendleton is shown on the chart as equivalent to both formations.

Percha shale.—Fauna is composed mostly of Upper Devonian types, but a few species such as *Athyris coloradoensis* suggest Mississippian types. The shale is dated by the similarity of *Syringospira prima* to *"Cyrtia" alta* of the Amity shale situated in the lower part of the Conewango stage.

The so-called Percha, Sly Gap shale (Stevenson, 1941, p. 163) of the San Andres and Sacramento Mountains of south-central New Mexico contains the Independence fauna and is older than the type Percha farther west in New Mexico near Silver City. It is interesting to note that the Sly Gap fauna includes *Spirifer orestes* along with more typical Independence species, and the presence in it of *Tropidoleptus carinatus* is a link to the Chemung of New York.

Petoskey formation (Cooper and Warthin).—Contains several faunas of different age. At the base is an assemblage suggestive of the Norway Point of Alpena County, Michigan. Above this is a fauna resembling that of the Potter Farm formation, exposed at the junction of U. S. highway 31 with Michigan State highway 131. This is followed by an assemblage containing an element of the Thunder Bay fauna, the blastoid *Heteroschisma gracilis* (Wachsmuth), and exposed on the Lake Michigan shore 1¼ miles north of Norwood.

Above the Thunder Bay equivalent occurs a fauna that must be assigned to the Upper Devonian (Sherburne). Here occur a few feet of thin-bedded limestone and soft shale (?) containing *Actinoptera, Schizophoria, Leptaena,* and large *Pentamerella.* On top of this zone an undetermined thickness of shale (not over 3 feet) contains nodose *Melocrinus, "Reticularia"* cf. *"R." laevis,* and *Pugnoides.* The presence of the last two places the upper fauna as of Sherburne (*"Reticularia laevis"*) age. The *"Reticularia"* bed is overlain by black shale and greenish-gray limestone containing *Styliolina* and *Buchiola* which are correlated with the Naples (Cashaqua).

Portland Point limestone.—Shell breccia and crinoidal limestone extending from Erie County to Potter Hollow, Albany County, New York. In Schoharie Valley the member is a quartz pebble band containing characteristic Portland Point fossils (such as *Centronella*). Along the Delaware River Cooper and Chadwick discovered the Portland Point with abundance of *Vitulina* and occasional *Centronella* in Elks Charles Brox Memorial Park and in a small glade ½ mile southeast of Sparrowbush near Port Jervis, New York. The member has not yet been seen in Pennsylvania but should occur there.

Potter Farm formation.—Contains species related to those of the Thunder Bay formation and elements of the Cedar Valley formation.

Ragland sandstone.—Contains *Tropidoleptus carinatus, "Spirifer" venustus, Leptaena,* and *Anoplotheca camilla* (reported as *A. acutiplicata*) associated with abundant corals and other common Hamilton types. The corals and *"S." venustus* suggest correlation with the Centerfield formation, but the presence of *A. camilla* and *Leptaena* suggest a greater age, possibly *Marcellus,* as both species are known that low in the Hamilton (Cleland, 1903, p. 20-21; Cooper, 1933, p. 549). The Ragland is therefore tentatively correlated with the *Meristella*-Coral zone of the Schoharie Valley sequence occurring about 250 feet above the Onondaga limestone and containing *"Spirifer" venustus.*

Rensselandia [*Newberria*] *zone.*—The brachiopod *Rensselandia* [*Newberria*] is a large, generally elongate, smooth-shelled brachiopod resembling *Amphigenia* or *Rensselaria.* In the United States five species are known: *R. claypolei* occurring in the Montebello sandstone of Pennsylvania and the Beauvais sandstone of Missouri; *R. johannis* Hall and *R. cordiforme* Stainbrook occurring in the *Profunda* zone of the Solon member of the Cedar Valley not far above *Hypothyridina intermedia; R. missouriensis* (Swallow) from a zone between the Cooper and Callaway now called the Ashland limestone; *R. laevis* (Meek) occurring with

Stringocephalus in the Mackenzie Valley, northwest Canada. A species not yet described is known from the Rogers City limestone of Presque Isle County, Michigan. In Germany the genus is known from the upper part of the Eifelian and from lower and upper Givetian; its range is thus the same as that of *Stringocephalus. Rensselandia* thus has particular significance in stratigraphy in indicating the *Stringocephalus* zone where that genus is absent.

In the United States and Canada *Rensselandia* is known with *Stringocephalus* in the Mackenzie Valley and occurs in the Rogers City with other *Stringocephalus* associates. In Iowa and Missouri it occurs in equivalents of the Tully, and in Pennsylvania it is known in a considerable thickness of middle Hamilton rocks. Consequently the Hamilton above the Marcellus plus the Tully are regarded as the equivalent of the Givetian.

Rogers City limestone.—Exposed in the great quarry of the Michigan Limestone and Chemical Company at Rogers City, Presque Isle County, Michigan. Consists of 70 feet of limestone overlying the Dundee limestone containing *"Spirifer" lucasensis*. The Rogers City contains a fauna including elements of the Elm Point, Winnepegosis, and Manitoba limestones such as *Büchelia tyrrelli* and *Omphalocirrus manitobensis* (Whiteaves). The brachiopod *Rensselandia* is also known and emphasizes the relationship to the *Stringocephalus* zone. Although *Stringocephalus* has not yet been found, the Rogers City molluscan fauna contains bizarre types known elsewhere on this continent only from the limestones of Manitoba (Ehlers and Radabaugh, 1938, p. 441-446).

St. Laurent limestone.—According to present knowledge this limestone occurs in two areas not yet satisfactorily correlated. One area is located in southwestern Illinois in Union and Jackson counties and an adjacent series of outcrops on the west side of the Mississippi in the Altenburg Quadrangle, Missouri. The other is in the Little Saline Valley south of Ste. Genevieve, Missouri.

Above the Grand Tower limestone in southern Jackson County occurs a zone about 10 feet thick containing the coral *Microcyclus discus*. Overlying this zone about 40 feet of limestone abound in *Tropidoleptus* and *Chonetes* cf. *C. coronatus*. The *Microcyclus* zone was seen in eastern Missouri at Union School about 0.8 mile south of Seventysix. Above this zone at the same locality occurs a considerable thickness of Hamilton rocks capped by a granular limestone abounding in corals and suggestive of the Lingle limestone (*Vitulina* zone).

In Little Saline Creek Valley and on St. Laurent Creek 100 to 275 feet of sandy limestone contains Hamilton fossils. No zones were recognized as common to the two localities, and none of the zones of the more southern Jackson County area, such as the *Microcyclus* zone, were recognized in the Little Saline Valley area.

The writer regards the St. Laurent limestone as lower Hamilton (Cazenovia stage) and below the Lingle limestone (*Vitulina* zone). The *Microcyclus* beds are regarded as probably of Marcellus age, but to them it may be necessary to add a part of the upper Grand Tower containing a *Spirifer* suggestive of *S. lucasensis* of the Dundee. At Devils Bakeoven and in the St. Laurent of the Little Saline Valley occurs a large pelecypod of the genus *Gosselettia* (*G. triquetra*) suggestive of the Marcellus and Skaneateles. The present position of the St. Laurent is tentative, but it seems unlikely that the formation can be placed above the Lingle as no post-Centerfield (Tioughnioga) formations are now known in the Mid west.

Schoharie formation in New York (Goldring and Flower).—The Schoharie formation in the Schoharie Valley region is typically a mud-textured siliceous limestone, dark bluish-gray when fresh, weathering to a brown porous sandrock, with an estimated thickness of 5 to 6 feet which dwindles in its most western exposure at Cobleskill to 35 inches of a fine-grained, siliceous limestone, with chert, characterized by a poor fauna of corals, few brachiopods, gastropods, and trilobites instead of the typical rich cephalopod fauna, with brachiopods, gastropods, and pelecypods abundant. There is no sharp contact between the Schoharie and Onondaga. Eastward in the Helderbergs (Capital District), where the typical Schoharie is best exposed, the formation shows an upper 32-inch heavy bed of dark blue-gray, fine-grained siliceous rock, weathering into a spongy sandrock, the portion that characterizes the Schoharie valley. The lower 30 inches is thinner-bedded and, when fresh, resembles a dark-blue, fine-grained limestone and appears to be the portion of the Schoharie that thickens southward in the Hudson Valley where the Schoharie formation loses the typical lithology entirely and becomes a shaly limestone with chert in the upper portion. This facies of the Schoharie, with a radical change in the composition of the fauna to one consisting largely of small brachiopod genera with a representation of a few bryozoans, corals, gastropods, cephalopods, and trilobites, attains a maximum thickness of 80 to 90 feet (12 to 17 feet transitional) in the Kingston region, previously interpreted as upper fossiliferous Esopus. These beds steadily thicken southwestward with much the same lithology and fossil content until in the Port Jervis region (Orange County) a thickness of between 215 and 235 feet is reached in beds in "Trilobite mountain" previously identified as basal Onondaga.

Schoharie fauna.—The work of Goldring and Flower has clarified one of New York's puzzling faunas. The supposed upper Esopus fauna consisting of a few species including *Eodevonaria, Leptocoelia, Anoplotheca,* and *Chonostrophia* now proves to be a facies of the Schoharie. The genera named suggest the Camden chert of Tennessee and its correlates in Illinois and Ontario. The fauna of the Camden and its correlates has been difficult to place because of the numerous Oriskany elements. It now assumes its proper position when correlated with the Schoharie. The so-called Onondaga of the Southern Appalachians now proves to be Schoharie, a part of the Onondaga rather than the whole formation.

Seneca limestone.—*See* Onondaga.

Sheffield formation.—Laudon (1935, p. 246) reports *Athyris angelica, Productella lachrymosa,* and *Spirifer disjunctus* and suggests a correlation with the New York Chemung. In New York these fossils overlie the restricted Chemung, and in Iowa the Sheffield overlies the restricted Chemung equivalent, Hackberry shale. It thus seems clear that the Sheffield is of Cassadaga age and not Kinderhook as claimed by some.

Shell Rock limestone.—Fauna is of early Upper Devonian age containing: *Pachyphyllum, Hypothyridina, Tenticospirifer* [*Trigonotreta shellrockensis* Belanski], *Platyschisma.* This fauna had hitherto been thought to be confined to Iowa, but the writer's studies indicate that it is widespread in western United States, its elements occurring in the Jefferson, Muddy Peak, lower Devils Gate, lower Minnewanka, and parts of the Sultan and Silverhorn limestones. Belanski's *Trigonotreta shellrockensis* is close to *Tenticospirifer utahensis,* and his *Spirifer cardinalis* is similar to *"S." argentarius.*

Silica shale.—That the Silica shale is referable to the Hamilton is a certainty, but its precise position is uncertain. According to the writer's views the Silica section covers a fairly long time interval ranging from upper Marcellus to high Skaneateles. The Silica shale overlies the "Blue-bed" of Carman (*in* Bassett 1935, p. 437-439) which is probably of high Dundee age and is overlain by the Tenmile Creek limestone of Centerfield age. The Silica thus occupies an interval between Marcellus and basal Ludlowville.

The "Blue-bed" contains the pelecypod *Gosselettia,* a good index to the lower Hamilton. Above this basal bed 14 feet of shale abounds in *Paraspirifer bownockeri, Mucrospirifer prolificus,* and *Chonetes* cf. *C. coronatus.* As *Paraspirifer* is abundant in the Mount Marion of New York and is found in the Solsville and Mottville members the Silica is correlated with the upper Marcellus. Incidentally the fauna of the

Mount Marion occurring at the top of Great Falls west of Catskill is very close to that of the Silica shale.

In the Silica Quarry the *Paraspirifer* zone is terminated for convenience at the *Rhipidomella* zone. In the section along Tenmile Creek between the *Rhipidomella* bed and the Tenmile Creek limestone occurs 6 feet of shale that contains *Schizophoria, Camarotoechia,* and other fossils suggestive of the Ferron Point and perhaps the Genshaw of the Michigan sequence.

Silver Creek limestone.—Rests on the Speeds limestone of upper Marcellus age and underlies the Centerfield equivalent, the Beechwood limestone. Although the Silver Creek contains elements of the Solsville sandstone fauna, it seems best correlated with the Delphi Station shale. In common with the latter it contains *Roemerella, Macrochilina, Bembexia, Limoptera,* and *Paracyclas. Conularia* and *Ptychodesma* suggest a Solsville age.

Silverhorn dolomite.—Gray dolomite capped by a thick bed of quartzite exposed in the Bristol Quadrangle and in the vicinity of Pioche, Nevada. *Stringocephalus* was found loose at Dutch John Mountain (Westgate and Knopf, 1932, p. 16-19), 40 miles north of Pioche, and its source was undoubtedly beds low in the Silverhorn. At 2300 feet up in the section south of Silverhorn *Tenticospirifer utahensis* was found. At other sections *"Spirifer" argentarius* occurs. It is evident therefore that part of the Silverhorn is referable to the *Stringocephalus* zone (upper Nevada limestone) and part to the *S. argentarius-T. utahensis* zone (upper 700 or 800 feet) which is correlated with the lower Devils Gate and Jefferson limestone.

Skunnemunk conglomerate.—Consists of conglomerates alternating with red sandstones, the formation varying in thickness from 300 to 2500 feet. Has yielded no fossils, consequently in order to arrive at an age assignment the Green Pond and Skunnemunk Mountain sections may be compared with other known areas of the Devonian. According to the writer's views this great outlier is a continuation of the sequence west of Catskill and represents about the same facies as found in that area. From Albany and Catskill the Hamilton outcrop belt extends seaward rather than parallel to the old shore. This is shown by the fact that along the Catskill Front the marine facies of the Hamilton comes more and more to dominate the section, until the entire sequence is marine at Port Jervis. The shoreward continuation of the Catskill sequence the writer believes to be the Skunnemunk outlier and its continuation into New Jersey.

The stratigraphic sequence of the outlier and the thicknesses involved are:

	Feet
Skunnemunk conglomerate	300-2500
Bellvale sandstone	1000-1800
Cornwall shale	200-1000

Willard (1937, p. 271-272) assigned the Cornwall shale to the Onondaga. This would leave 1300-4300 feet of post-Onondaga sediments. Willard assigned the Bellvale to the Marcellus, and this seems correct when one considers that about 1300 feet of Marcellus is known in the vicinity of Catskill (Bakoven 180 feet, Mt. Marion 800 feet, and Ashokan 300 feet). This leaves the remaining 300 to 2500 feet equal to the remainder of the Hamilton which in Schoharie Valley is about 1800 feet thick and is much thicker on the Catskill Front although the limits are not exactly known (probably about 3000 feet). On this evidence the entire sequence, even allowing for part of the Cornwall to be Hamilton, is probably not younger than Hamilton. A possible Taghanic equivalent could be present, but the writer favors the view that the entire sequence is Hamilton and Onondaga only and has so represented it on the chart.

Sly Gap formation.—Named by Stevenson (1941) for dark shale and limestone outcropping in the Sierra Caballo, and San Andres and Sacramento mountains, south-central New Mexico. Contains the Independence fauna according to Stainbrook (1935b). Stevenson correlates the Sly Gap with the Martin limestone of Arizona. Possibly partial equivalence with the Canutillo formation is suggested above. (*See* Canutillo, Percha shale.)

Snyder Creek shale.—Its generic assemblage is predominantly Middle Devonian, but a few Upper Devonian types are present: *Gruenwaldtia gregeri, Productella callawayensis, Strophonelloides crassus,* and *Dactylocrinus concavus.* Along with these occurs a species of *Pentamerella,* a genus that is much more abundant in the Middle Devonian. Because of some Middle Devonian aspects of the fauna and because it is definitely post-Tully the writer has placed it in the Finger Lakes (Genesee group) or lowest division of the Upper Devonian; possibly it should be placed higher with the Ithaca fauna.

Speeds limestone.—Name applied to light-brown limestone containing *"Spirifer" mucronatus* in southern Indiana (Sutton and Sutton, 1937, p. 331). Includes the *Hadrophyllum* beds of the Delaware and underlies the Deputy limestone.

Squaw Bay formation (Warthin and Cooper).—Dolomite with thin layers of limestone aggregating 3-12 feet in thickness. Contains a thin zone near the bottom nearly completely composed of shells of *Styliolina fissurella.* Other fossils are: *Paracardium, Buchiola, Diaphorostoma pugnus,* and the goniatite *Koenenites cooperi.* Correlated with the Genundewa of New York on the basis of abundance of *Styliolina,* the *D. pugnus,* and general assemblage. Regarded as lowest Upper Devonian of eastern United States because of presence of *Koenenites* which occurs in the lowest zone of Upper Devonian goniatites in Germany.

Starbird formation.—Named from Starbird Ridge, Windermere, Kootenay District, British Columbia. Consists of arenaceous and argillaceous limestone about 230 feet thick on the east end of the ridge. Kindle (*in* Walker, 1926, p. 35) reports *"Spirifer"* cf. *"S." argentarius, "S."* cf. *"S." raymondi,* and *Schizophoria* which suggest an Upper Devonian age and correlation with the Jefferson and lower Minnewanka limestones.

State Quarry limestone.—Occupies erosion pockets in the Cedar Valley in a limited area in east-central Iowa. Fauna includes abundance of *Pugnoides, Atrypa,* and *Athyris,* and a species of nodose *Melocrinus* (Stainbrook and Ladd, 1926). The last mentioned and the *Pugnoides* suggest a position low in the Upper Devonian, and the formation has therefore been placed at the top of the Genesee group. Possibly it should be placed higher in the Naples stage (Ithaca).

Stone Mill limestone.—A lens of crinoidal limestone containing the Centerfield fauna located on both sides of Chenango Valley south of Hamilton, New York. Contains a new species of *Prismatophyllum.* The Ludlowville shaly sandstone overlying the Stone Mill also contains the Centerfield fauna. The Stone Mill represents the last recognizable phase of the Centerfield where it passes into the undifferentiated sands east of Chenango Valley.

Stony Hollow sandstone.—A conspicuous layer of calcareous sandstone 75-100 feet thick exposed at Port Jervis, and from Kripplebush northeastward to the southwest corner of Albany County, New York. Known also near Echo Lake, Pennsylvania. Between Kripplebush and Catskill it forms a conspicuous line of cliffs and waterfalls. Northeast of Catskill this sandstone has been traced to the fall of Onesquethaw Creek and from there westward where it becomes the Cherry Valley limestone in Schoharie Valley. Fauna contains many new species, among them a new *Pentamerella* that also occurs in the Cherry Valley limestone and a species of the rare brachiopod *Kayserella.* Tracing of the Stony Hollow into the Cherry Valley has brought order to the sequence along the Catskill Front; it has proved that the Bakoven is equivalent to the Union Springs member and that much of the Mount Marion belongs to the Chittenango member.

Stringocephalus zone.—Because it is large and easily identified and has a fairly restricted range, *Stringocephalus* is an important index fossil. In Germany where it is most abundant, its range is from the upper Eifelian through the Givetian, and in general practice its presence indicates the upper half of the Middle Devonian (Givetian). In the United States *Stringocephalus* occurs in southeastern and central Nevada, in western Utah, and southeastern Alaska. In Canada it has been reported from the Ramparts of the McKenzie River, Great Slave Lake, and Lake Winnepegosis, Manitoba (Kindle, 1921, p. 21-24; Kirk, 1927, p. 219-222). In Midwestern United States the genus has been reported from southern Minnesota (Stauffer, 1922, p. 408), but the discovery has not been authenticated. In Presque Isle County, Michigan, many species of the Manitoba *Stringocephalus* limestone have been taken from the Rogers City limestone, which lies between the Marcellus (Dundee) limestone and Traverse (Bell) shale.

The position of *Stringocephalus* between the Dundee and the Bell places the remainder of the Hamilton (Skaneateles to Moscow) and the Tully in a post-*Stringocephalus* position. But the faunas of these formations indicate a Middle rather than Upper Devonian age. Consequently it is believed that the Hamilton (above the Marcellus) and the Taghanic belong in the upper part of the *Stringocephalus* zone, corresponding to the upper *Stringocephalus* stage of Germany.

Although it is clear that the Rogers City limestone is low in the *Stringocephalus* zone it is more difficult to place the *Stringocephalus* beds of the West. In the Eureka district *Stringocephalus* overlies a Marcellus equivalent and is overlain by the *Stromatopora* zone, followed by the lower Devils Gate formation of lower Upper Devonian age. Perhaps the *Stringocephalus* zone of the West equals the Rogers City plus the Hamilton and the Tully, but no fossils accompanying *Stringocephalus* are known to prove this point.

Little help in this dilemma can be got from the fauna of the *Stringocephalus* zone at the Ramparts of the McKenzie River, a fauna that contains elements of the Nevada limestone (*Leiorhynchus castanea* Meek) and elements of the Winnepegosis limestone (*Rensselandia* and *Gypidula*). This evidence suggests that the American and Canadian occurrences represent a single zone. The writer knows no contradictory evidence and has therefore placed all the *Stringocephalus* occurrences as equivalent and low in the Givetian. The longer range of *Rensselandia* is taken to show the upper Givetian or upper *Stringocephalus* age of the upper Hamilton, Cedar Valley, upper Traverse, and Tully.

Tenmile Creek limestone.—Overlies the upper Silica shale on Tenmile Creek, about a mile south of the Silica Quarry, southwest of Sylvania, Ohio. Contains the Centerfield species *Strombodes alpenensis* and *Cyclorhina nobilis* associated with many species of Centerfield type.

Threeforks shale.—Sequence and faunas well described by Haynes (1916, p. 13-30) who reports about 145 feet of shales and beds of limestone. Above this sequence occurs yellowish sandy limestone containing *Syringothyris*. In the Threeforks proper occur a variety of fossils including large *Cyrtospirifer, Productella,* and *Leiorhynchus* suggesting affinity with the Percha shale of New Mexico and the Ouray limestone of Colorado. The pelecypods and goniatites on the other hand indicate a somewhat older age. Of the former the most important is *Loxopteria* which in New York is restricted to the Canadaway group. The goniatites give another clue to the position of the Threeforks by making comparison with the German column possible. The most important Threeforks goniatites are: *Raymondiceras simplex* (Raymond), *Platyclymenia* (*Pleuroclymenia*) *americana* Raymond, and *P.* (*P.*) *polypleura* Raymond which, according to Schindewolf (1934), indicate correlation with the *Prolobites-Platyclymenia* stage (Oberdevon III). As this stage overlies the *Manticoceras* zone (Genesee to Canadaway groups), the Threeforks is placed in the succeeding Conneaut group of the Cassadaga stage. This position is corroborated by the occurrence in the Threeforks of brachiopods almost identical to *"Camarotoechia" duplicata,* chief index to the Conneaut.

The *Syringothyris* beds above the Threeforks and below the Mississippian (Madison) formation Schuchert (1910, p. 546) assigned to the Louisiana limestone. As *Syringothyris* is known in the Conewango stage (Chagrin) and as the Mississippian age of the Louisiana has been questioned, it is not unlikely that the Threeforks *Syringothyris* zone will ultimately be assigned definitely to the Devonian (Conewango). For the time being the beds are placed in the Devonian or Mississippian column.

In northeastern Utah a zone of dark shaly rock between the Jefferson and the Madison has been referred to the Threeforks. This zone contains large *Schizophoria,* large *Productella?, Syringothyris, Rhipidomella,* and a *Spirifer* suggesting *S. marionensis* of the Louisiana. This shale is also placed in the Devonian or Mississippian column as it is definitely not Threeforks shale.

Thunder Bay formation (Cooper and Warthin).—This name preoccupies Partridge Point formation of Warthin and Cooper (1935, p. 525). The fauna contains Cedar Valley as well as Hamilton elements such as: *Nucleocrinus obovatus* Barris, *Heteroschisma gracilis* (Wachsmuth), *Strobilocystites?*, and *Stereocrinus*. The brachiopods contain Cedar Valley and Hamilton types, but the pelecypods are mainly New York Hamilton forms. Another New York Hamilton link is the trilobite *Dipleura dekayi*. As this formation underlies the Squaw Bay limestone, a correlate of the Genundewa limestone, it could represent a calcareous facies of the Geneseo, but the writers prefer to correlate it with the Tully because of the close faunal tie with that formation and the Cedar Valley. Furthermore the beds assigned to the Thunder Bay on the shore of Lake Michigan contain a trilobite identified with *Scutellum tullium depressum* Cooper and Cloud (1938, p. 459) which occurs in the lower Cedar Valley of Calhoun County, Illinois.

Tully formation.—For many years the Tully had been placed as the top member of the Hamilton group, but in 1890 H. S. Williams (p. 481-500) presented evidence designed to show that this limestone is the homotaxial equivalent of the Upper Devonian *"Cuboides"* zone of Europe. Re-examination of the evidence and new information indicate that the Tully may, rather, be the homotaxial equivalent of the upper *Stringocephalus* zone (Wells, 1940) and that the true homotaxial equivalent of the European *"Cuboides"* zone is the Independence-Nunda (High Point) fauna high above the Tully.

In discussing the Tully fauna H. S. Williams neither considered nor understood the entire fauna. His collections, which are in the U. S. National Museum, were taken from the lower part of the Tully only (Tinkers Falls and Apulia members) at Tinkers Falls, Tully, Cuyler, and a few neighboring localities. The collection thus does not embrace all the Tully of the State nor does it include the fauna of the West Brook member of Cooper and J. S. Williams, which was well known to S. G. Williams (1887, p. 26-28). This latter fauna H. S. Williams (1890, p. 496, footnote) erroneously said was derived from Hamilton shales below the Tully. H. S. Williams further regarded the Tully fauna as consisting of about 50 species, but the formation is now known to contain more than 260 species most of them Hamilton forms or forms with Hamilton antecedents. It is therefore evident that H. S. Williams had no conception of the total Tully fauna.

According to H. S. Williams the more important Tully species are: *Schizophoria tulliensis, Schuchertella arctostriata, Leptostrophia tulliensis, Chonetes aurora, Atrypa reticularis, A. aspera, Hypothyridina* *venustula, Spirifer mucronatus tulliensis, Cyrtina hamiltonensis, Spirifer tullius, Ambocoelia umbonata, Productella tulliensis, Elytha fimbriata, Phacops rana, Greenops boothi, Scutellum tullium, Platyceras symmetricum.* Eleven of these species—2, 5, 6, 9-15, and 17 of the list—can be dismissed from consideration in age determination because they are long-ranging Hamilton types. This leaves a few exotic forms for consideration. *Schizophoria tulliensis* now has little significance because the genus is well known in the lower Hamilton from which many Tully species were derived.[2] *Leptostrophia tulliensis* shows characters that link it with the *Nervostrophia* stock of the Ithaca rather than with Hamilton forms. *Chonetes aurora* may be related to Hamilton species as the genus is abundant there, and the species is identified from Manitoba. A nearly identical species occurs in the Solon of Iowa. *Hypothyridina venustula* is an exotic form, but its antecedents are known in the pre-Tully Eifelian and Givetian. *Spirifer mucronatus tulliensis* is related to *S. mucronatus* and may represent passage to the Ithaca *S. posterus* by greater development of the dental plates. *Scutellum tullium* has Middle Devonian antecedents in the Givetian of Europe and the Winnepegosis limestone of Manitoba. Aside from *Nervostrophia, Spirifer mucronatus tulliensis,* and a possible representative of the genus *Manticoceras* (Miller, 1938, p. 74-76), no unequivocal Upper Devonian types occur in the Tully. If the Tully actually were homotaxial with the *"Cuboides"* zone of Europe more than two or three exotic types would be expected. In this respect the Tully fauna cannot be compared with that of the Independence-Nunda in which many European types appear in a resident fauna.

That the Tully cannot be the homotaxial equivalent of the European *"Cuboides"* zone is shown by the fact that, besides *Hypothyridina* and *Scutellum,* no other elements of the *"Cuboides"* fauna are found in it. One of the important *"Cuboides"* types is *Cyrtospirifer verneuili,* but this genus in North America has not been seen below the Naples. Other important *"Cuboides"* genera are *Gypidula, Ambothyris,* and *Pugnoides,* but these have never been seen in the Tully. *Gypidula* occurs in the Cedar Valley, a Tully correlate, yet there it occurs with *Rensselandia* and belongs to the Middle Devonian. Thus few European elements appear in the Tully, but the Independence fauna or *"Cuboides"* equivalent contains many European types including *Cyrtospirifer.* It seems best now to regard the Tully as of Middle Devonian age, and to make a stage, the Taghanic stage, to include it and its correlates in the Midwest and the overlying Geneseo shale which contains *Hypothyridina* in Kentucky.

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[2] After disappearing before the Ludlowville and Moscow, *Bembexia sulcomarginata* and *Paracyclas lirata,* both Marcellus and Skaneateles types, reappear in the Tully.

Valentine limestone.—Contains *Platyschisma mccoyi* and a small spiriferoid similar to *Tenticospirifer utahensis* indicating correlation with the lower Devils Gate and the Jefferson limestone.

Wapsipinicon limestone.—Poorly fossiliferous sediments overlying the Niagaran and underlying Cedar Valley limestone. According to Stainbrook (1935a, p. 249-252) consists of the following members in ascending order: Coggon dolomite member, Otis limestone, Kenwood shale, Spring Grove limestone, and Davenport limestone. The Coggon contains a brachiopod suggesting *Emanuella subumbona* but specifically different, and a small *Conocardium.* The Otis contains *Emanuella,* but the rest of the members are without fossils. The Wapsipinicon is referred to the Taghanic stage because of the similarity with the Cooper limestone and Potter Farm formations. The Cooper contains a small *Emanuella,* and a small *Conocardium* is abundant in the Potter Farm. The latter also contains a snail suggestive of *Turbonopsis providencis* of the Cooper.

West Range limestone.—Contains *Athyris "angelica,"* rotund *Cyrtospirifer, Pugnoides,* and a fine-ribbed *Camarotoechia* and is placed in the Canadaway.

In the Dutch John Mountain section 40 miles north of Pioche Merriam (1940, p. 39) shows that *Cyrtospirifer* ranges for 1835 feet and that the limestone is divisible into five zones. The two lowest, A and B, contain *Cyrtospirifer* and *Pachyphyllum* and are referred to the Chemung. Zone C, or *Cladopora* zone, follows, and on this is zone D which abounds in brachiopods of the West Range formation and is assigned to the Canadaway group of the Cassadaga stage. A fifth zone (E) is not placed.

Williams Island limestone.—In James Bay region includes upper 8 feet of Abitibi River limestone of Savage and Van Tuyl and 30 feet of drab, soft limestone above it (Kindle, 1924, p. 34). The section is capped by 2 feet of limestone containing *Hypothyridina.* According to Kindle the rock below *Hypothyridina* contains Hamilton fossils. Therefore the part containing these fossils has been placed in the Hamilton, and the *Hypothyridina* bed assigned to the Tully.

Winnepegosis limestone.—*Stringocephalus*-bearing limestone related to the Elm Point and Manitoba limestones. (*See* Manitoba.)

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